WATERSPOUT OBSERVED OFF THE COAST OF SOUTHERN SICILY IN AUGUST, 1876.

Frontispiece.

A POPULAR TREATISE

ON THE

WINDS:

COMPRISING THE

GENERAL MOTIONS OF THE ATMOSPHERE, MONSOONS, CYCLONES, TORNADOES, WATERSPOUTS, HAIL-STORMS, ETC. ETC.

BY

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PREFACE.

SINCE the middle of the present century great advances have been made in meteorology, especially in the study of the mechanics of the atmosphere. Before this epoch little was known even with regard to the general motions of the atmosphere, the true theories of cyclones, tornadoes, water-spouts, hail-storms, cloud-bursts, etc., were entirely unknown, and the observed phenomena were mostly regarded as mysteries. Although there are still some things which require more study and further explanation, yet these subjects have now become much clearer and better understood; and so great has the change been, that the recent advances are often called the "new meteorology."

During this period of advancement the writer has had published a number of meteorological papers in an endeavor to advance our knowledge in the subjects mentioned above. The first of these, entitled "An Essay on the Winds and the Currents of the Ocean," was published in the Nashville Journal of Medicine and Surgery in the year 1856. The writer's attention was first directed to this subject by reading Maury's Physical Geography of the Sea, from which he learned that the pressure of the atmosphere is less both at the poles and at the equator of the earth than it is over two belts extending around the globe about the parallels of 30° north and south of the equator. An attempt to account for this phenomenon, which was then inexplicable upon any known principles, resulted in the essay named above. This essay was mostly of a popular character, and only a very imperfect, though at the time important, first step, which subsequently led to further researches, and the discovery of the effect of the earth's rotation in the dynamics of the atmosphere, which has served to clear up many of the

mysteries of the older meteorology. This essay was subsequently republished by the Signal Service in Professional Paper of the Signal Service, No. XII.

In the year 1858 this subject was again taken up by the writer, but treated in a very mathematical manner, and the results were given in a paper entitled "The Motions of Fluids and Solids relative to the Earth's Surface," which was published in a series of parts in Runkle's Mathematical Monthly, and subsequently republished by the Signal Service in Professional Paper of the Signal Service, No. VIII, with extensive notes, giving the mathematical processes in detail, by Professor Frank Waldo, then in the Signal Service.

This subject was taken up a second time by the writer a number of years afterward, but still treated in a very mathematical manner, and the researches so extended as to embrace cyclones, tornadoes, water-spouts, hail-storms, cloud-bursts, etc. This resulted in his "Meteorological Researches for the Use of the Coast Pilot," published in three parts in the Reports of the Superintendent of the Coast and Geodetic Survey for the years 1875, 1878, and 1881.

In his "Recent Advances in Meteorology," published as the second part of the Chief Signal Officer's Report for 1885, the writer went over the same ground again, but adopted mathematical methods somewhat simplified and gave further extensions of the more popular parts of the subjects.

Although these papers have been pretty widely distributed as Government publications, yet they have fallen into the hands of comparatively few of the large number of readers who are interested in the important subjects of which they treat, and they are also too mathematical, for the most part, for many readers. On account of the great and pretty general interest which has been taken of late years in meteorological subjects, there are now many who wish to obtain at least a general, if not a thorough, knowledge of the most important principles of these subjects, but who, either from lack of ability or time and inclination to take hold of difficult reading, are desirous of having some more popular presentation of them,

and the desire has often been expressed that the writer would undertake such a presentation of these subjects, treated mostly in a mathematical way in the papers referred to. To comply with this desire, so far as the nature of the subjects treated will admit, is the object of the present work.

From the nature of the subjects, however, little can be done where even the simpler operations of mathematics are entirely discarded, and the results of such a method of treatment would be very unsatisfactory to a large class of readers who have some knowledge of the simpler principles and operations of mathematics, so that these have been used in the work; but all that could well be done has been done in the way of mere verbal explanations and illustrations of the subjects. It is therefore thought that the work can be read with profit by those who are deficient in even the elementary principles of mathematics.

In the present more popular presentation of the subjects treated in the works referred to above, of course the methods are very different, and the whole matter is very much expanded, so that all has been rewritten except a few pages in the chapter on Tornadoes, which have been copied with little or no change from my "Recent Advances in Meteorology." In the mathematical treatment of the subjects, where this has been introduced, no claim is made to elegance of methods, and sometimes simple arithmetical methods are used instead of algebraic methods and the calculus, such as almost any one well versed in the higher methods of analysis frequently adopts in the first discovery of important principles and results, or which he may use as a check of results obtained in a more mathematical way, the principal object being to give some insight into the subjects, though in a homely way, to readers who have little facility in the use of mathematics.

The subject-matter contained in the following pages is mostly an expansion of a series of forty lectures delivered by the writer before a class of army officers of the Signal Service during the months of February and March, 1886, in which the manner of presentation of the matter was somewhat the same

as that adopted here. This was necessary, since the members of the class had no time for reading and study, and so for entering into the matter more thoroughly, for all their time not given to the hearing of the lectures had to be given to their daily work in the service.

A thorough knowledge of the principles contained in the following work, it is thought, will be of great advantage to all who are desirous of understanding the observed phenomena and sequences of the weather, and of forming rules for weather prediction, and especially to the seaman on the ocean in the management of his vessel. For although no attempt has been made to lay down detailed rules to be followed in the various cases, yet the principles are given and explained by which each one with a knowledge of them can intelligently form his own rule in any case, which is generally much better than to blindly follow rules which can rarely be given so as to cover all cases without exceptions.

So much interest is now being taken in meteorology that its principles must soon be taught to a considerable extent in our colleges and universities, and it must form a part of any course in Physics where a separate chair is not assigned to it. It is therefore thought that lecturers on meteorological subjects before college classes, or any general and larger audiences, will find much in the following work which can be advantageously and conveniently used.

The author is indebted to the courtesy of the Chief Signal Officer for permission to copy from the publications of the Signal Service a few of the illustrations contained in the following pages.

He is also indebted to the kindness of his friend Professor Frank Waldo for the full and well-arranged index to this work.

WM. FERREL.

KANSAS CITY, MO., April, 1889.

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CHAPTER I.

THE CONSTITUTION AND NATURE OF THE ATMOSPHERE.

COMPOSITION OF THE ATMOSPHERE.

1. THE gaseous envelope surrounding the earth called air, and regarded as a whole, the Atmosphere, is composed of the two principal gases in its constitution, oxygen and nitrogen, together with a small amount of carbonic acid and traces of other gases. These constitute what is called *dry air*. When composed of oxygen and nitrogen only, it is said to be *dry* and *pure*. In addition to the constituents of dry air the atmosphere contains also a variable amount of aqueous vapor, which, in a warm climate, may amount to one-twentieth part, or more, of the whole. The proportions of dry air, neglecting the slight and mostly unmeasurable traces of the other constituents, may be put as follows: Oxygen, 20.95; nitrogen, 79.02; and carbonic acid, 0.03 parts, in 100 by volume.[1] *

The constituents of the atmosphere are not chemically combined, but exist together as a mechanical mixture. If the atmosphere were perfectly at rest each constituent would assume the same status and distribution as it would if the others were not present. Each one, therefore, would form an atmosphere of itself, independent of, and unaffected by, the others. This arrangement of the constituents is called *Dalton's law*. But as perfect quiet for a long time is a condition of this arrangement, on account of the almost continual agitations of the atmosphere, this law is never satisfied.

* For the references see the Appendix.

BOYLE AND MARIOTTE'S LAW.

2. Gases are extremely elastic, so that a given amount of any one of them may be compressed into a space almost infinitely small; and if the pressure is sufficiently removed, it expands into a space almost infinitely great. The space which a given amount of gas occupies under varying pressures and temperatures is its volume; and the space which a unit mass, as one gram or one grain, occupies under an assumed standard pressure and temperature is called the *specific volume.* The standard pressure is that of the mercurial column of the barometer 760^{mm} in height at the temperature of $0°$ C., and subject to the action of the force of gravity at sea-level on the parallel of $45°$; and the standard temperature is that of melting ice.

If we let P represent the pressure and V the volume of any given amount of gas at any temperature, then, however much the pressure may vary, the temperature remaining the same, we have sensibly the product PV equal to a constant. The law represented by this expression is called *Boyle and Mariotte's law,* from the names of its two independent discoverers, of whom Boyle seems to have been the first.

If we let P_0 represent the standard pressure, as defined above, and V_0 the specific volume of the gas, we have, whatever may be the pressure,

$$PV = P_0 V_0,$$

in which V is the volume of unit mass at standard temperature of 0°C. when subject to the pressure P.

According to Boyle and Mariotte's law, whatever the temperature of the gas, if P is doubled V is reduced to one half, or if P is diminished to one half V is then doubled, and so for any other ratios, either integral or fractional; and, consequently, if P is infinitely great, V becomes infinitely small, and *vice versa.*

The elastic or expansive force of a gas, which in a static state of the gas is exactly equal to the pressure, arises, according to the kinetic theory of gases, from the action of the atoms or molecules upon one another and upon the sides of the containing vessel in their numerous contacts in flying with very

great and differing velocities in all directions. The molecules of different gases at the same temperature have, on the average, different velocities: and as the forces are as the squares of the velocities, different gases have different elastic forces and volumes under the same conditions of pressure and temperature.

3. The law of Boyle and Mariotte is not strictly true for any of the gases, or for the atmosphere regarded as a simple gas; and this is especially the case where the pressure is either extremely great or extremely small. According to Regnault's experiments, the product PV in the case of the atmosphere decreases with increase of pressure up to the pressure of about 65 atmospheres, after which it increases, and for very great pressures it becomes much greater than it is in the case of the ordinary pressure of the atmosphere at the earth's surface. According to the experiments of Mendeleef[1] of St. Petersburg, the atmosphere conforms to Boyle's law at a pressure of about 650 millimeters of mercury; but for smaller pressures the product PV diminishes, and at very low pressures the air seems to almost entirely lose its elasticity and become an exceedingly rare liquid.

THE LAW OF CHARLES AND GAY-LUSSAC.

4. The elastic force of a gas and of the atmosphere is increased with increase of temperature. It is found from experiment that its volume increases the $\frac{1}{273}$ part of what it is at the temperature of melting ice for each degree Centigrade of increase of temperature. This is the *law of Charles and Gay-Lussac*, so called from its discoverers, of whom Charles is thought to have been the first. By this law the volume of unit mass of gas under the standard pressure P_0 and temperature τ becomes $V_0\,(1 + \frac{1}{273}\tau)$. We therefore have, as an expression of both the laws of Boyle and Charles combined,

$$PV = P_0 V_0\,(1 + \tfrac{1}{273}\tau),$$

in which V, for any given pressure P, is the volume of unit mass for the temperature τ.

By the law of Charles, as represented in the last member of this equation, a gas would lose all of its elastic force when cooled down to 273° below the zero of the Centigrade scale, since with $\tau = -273°$ the second member of this expression would vanish. As the elastic force depends upon temperature, or by the kinetic theory of heat, upon the velocities of the molecules in their impacts upon one another, where elasticity ceases temperature also must cease, and hence at 273° below the ordinary zero there cannot be any temperature. This point, therefore, is called *the absolute zero*, and the temperature reckoned from this point is called *the absolute temperature*.

If we put T for the absolute temperature, the preceding expression of the combined laws of Boyle and Charles in a function of T becomes

$$PV = RT,$$

in which $R = \frac{1}{273} P_0 V_0$. From this it seems that the pressure, if the volume remains constant, and conversely the volume, if the pressure remains constant, is proportional to the absolute temperature. Since different gases under the same temperature have different elastic forces, and consequently have different specific volumes, it follows that V_0 differs in different gases, and consequently the value of R in the expression above.

Since the pressure, and consequently the elastic force, in a static condition of the gas, is proportional to T when the volume is constant, and the elastic force depends upon and is proportional to the square of the velocities of the atoms or molecules, it follows that in the increase of velocities with the increase of temperature the mean square of the velocities is as the absolute temperature.

The law of Charles, as that of Boyle, is not strictly correct in all cases, but the deviations from the law within the range of experiments are extremely small.

ARRANGEMENT OF THE CONSTITUENTS.

5. If each of the several constituents of the atmosphere formed an atmosphere of itself around the globe, the arrange-

ments of the particles with regard to altitude would be different in the several gases, so that the densities would not decrease with increase of altitude in the same ratio. Since different gases have different elastic forces, the same weight of each under the same pressure occupies different spaces, or, in other words, has different volumes; and in order to have the same weights in a given space it is necessary for the gases to be subject to different pressures. As the density under the same pressure must be inversely as the volume where the temperature remains the same, the greater the elastic force of a gas the less its density. Hence the greater the elastic force of a gas forming an atmosphere around the globe, the higher it would be necessary to ascend to get above a certain part of it, as one half or any other proportion, and the more the whole mass would be expanded upward, where it is subject to the pressure only arising from the action of the force of gravitation upon its own mass.

In the case of a gaseous atmosphere surrounding the earth in a state of static equilibrium, the pressure at any altitude depends upon, and is very nearly proportional to, the mass of gas above that altitude, since the force of gravity of unit mass differs but little at the different altitudes in the atmosphere. Hence the density, since by Boyle's law it is as the pressure, decreases with increase of altitude; for the greater the altitude, the less the pressure of the part above that altitude. The greater the altitude, therefore, the greater must be the increase of altitude required to cause a given ratio of decrease of pressure of the whole, and of the density at the earth's surface. For instance, at the altitude where the pressure is reduced to one half of what it is at the earth's surface, it would be necessary to ascend twice as far to arrive where the pressure and density are diminished, say the $\frac{1}{100}$ part of what they are at the earth's surface, as it would at the earth's surface. If, therefore, Boyle's law held for infinitely small pressures, there would be no exterior limit to such a gaseous atmosphere, though it would become extremely rare at no very great altitude. But if, as may be the case according to

the experiments of Mendeleef, § 3, a gas under a very low pressure loses its elastic force and becomes a very rare liquid, then there must be a limit to such an atmosphere, but the density at this limit must be very small.

6. According to Dalton's law, deduced from experiments, and being also in accordance with the kinetic theory of gases, each constituent of our atmosphere in a perfectly quiet state would form a separate and independent atmosphere, with exactly the same arrangement of its parts, and the same relation of the densities of these parts with reference to altitude, as if the other constituents were not present. The constituents, therefore, with the greatest elastic forces and least densities under the same pressure would be expanded upward the most, so that whatever might be the relative densities of the independent gaseous atmospheres at the earth's surface, the densities of the more elastic gases would decrease with the increase of elevation at a less rate than those of the gases of less elastic force. For instance, if an atmosphere of hydrogen existed with one of nitrogen, the density of the latter under the same pressure and temperature being fourteen times greater, and consequently its elastic force as many times less, if the densities of both gases at the earth's surface were the same, at only a small altitude the density of the hydrogen would be much greater than that of the nitrogen, since it decreases with increase of altitude at a much less rate, and at an altitude where the nitrogen would sensibly vanish the density of the hydrogen would be diminished by only a small part; and if at the earth's surface the density of the hydrogen were very small in comparison with that of the nitrogen, yet at only a very moderate altitude it would be the predominating constituent.

7. The density of a gas, in relation to that of dry and pure air of the same temperature and subject to the same pressure, is called its *relative density*. The relative densities of oxygen, nitrogen, and carbonic acid gas are respectively 1.1056, 0.9714, and 1.52. The densities and consequently the elastic forces of the two principal constituents of our atmosphere being so

nearly the same, the relations between the densities of the two at the earth's surface and up at considerable altitudes, from what has been shown, would not be very different if the two existed together in accordance with Dalton's law. For instance, the ratios by volume of the two constituents, oxygen and nitrogen, at the earth's surface being as 20.95 to 79.02, respectively, § 1, there would be very nearly the same ratio up at a considerable altitude, the proportion of oxygen to that of nitrogen decreasing very slowly. The density of carbonic acid gas being much greater, and consequently its elastic force much less, its density decreases with increase of altitude at a much more rapid rate than that of oxygen and nitrogen, and consequently in the upper part of the atmosphere the proportion of this gas relative to that of the two others would be much less by Dalton's law than it is at the earth's surface; but this, we have seen, § 1, forms only a very small part of our atmosphere. This compound atmosphere, therefore, can be regarded as a simple gas; for although the several gases of which it is composed would not have, if it were in a quiescent state, quite the same ratios between them at the earth's surface and in the higher altitudes, yet on account of the continued agitation of the atmosphere in its general circulation and by storms, these relations are so nearly the same at different altitudes, that analyses of air at sea-level and on the tops of the highest mountains, and of samples obtained from very high altitudes in balloon ascensions, show no sensible differences in the proportions of oxygen and nitrogen. And as the rate of expansion with increase of temperature is sensibly the same for all gases, the laws of both Boyle and Charles can, therefore, be applied with sensibly the same accuracy as in the case of any simple gas.

8. According to Boyle's law, the atmosphere with constant temperature would occupy precisely the same volume, however much its mass were increased or diminished. If the mass were doubled or diminished to one half, the pressure, and consequently the density, would be changed in the same ratios, and so for any other proportions; and hence the volume would, in

all cases, be the same. The densities at the same altitudes, however much the whole mass of the atmosphere were increased or diminished, would have the same ratios to the densities at the surface, and it would in all cases be necessary to ascend to the same altitude to get above any given proportion of the whole. If, therefore, the atmosphere were of uniform temperature at all altitudes and the part above any given level, as that of the top of Pike's Peak, were considered by itself, it would form an atmosphere exactly similar to the whole, but the pressures and densities at equal altitudes above the base would be less, and in the same proportion as the mass of atmosphere above that level is less than that of the whole.

Since the densities at different altitudes in an atmosphere of uniform temperature decrease in the same ratio as the pressures in ascending vertically, and these, as the amounts of atmosphere remaining above, it follows that, at all altitudes, it is necessary to ascend through the same vertical distance, in order to get above a given proportion of the whole existing above the level of the starting point. For instance, if it is necessary to ascend from sea-level to the altitude of Pike's Peak to get above two fifths of the whole atmosphere, then it is necessary to ascend just as much higher to get above two fifths of what still remains above that level, and so on. Again, if it is necessary to ascend vertically a given distance to get above one tenth of the atmosphere, then it is necessary to ascend the same distance to get above one tenth of what is left; and so on, *ad infinitum*. After the first ascent there would be nine tenths of the whole mass left above, and consequently the pressure and density would be diminished to nine tenths; after the second ascent, to nine tenths of nine tenths of what they were at the earth's surface; and after the third ascent, in the ratio of the third power of nine tenths, or in the ratio of 1000 to 729, and so on. Hence if different altitudes are taken in arithmetical progression, the corresponding pressures and densities diminish in geometrical progression. And this is true whether we start on the earth's surface or at any level above

it, since, as we have seen, the part above that level forms an atmosphere precisely similar to the whole.

Where the temperature diminishes with increase of altitude, as in the real case of nature, the rate of increase of pressure and of density with increase of altitude, relatively to the whole, becomes greater above than below, and it is not necessary at any given level to ascend through so great distances, to get above a given proportion of the whole above that level, as it is at the earth's surface.

PRESSURE OF THE ATMOSPHERE.

9. The pressure of a body arises from the action of a force upon that body when it is not free to move, and this pressure is measured by the product of the mass of the body into the velocity which would be generated in any assumed unit of time, as one second, if the body were free to move in the direction in which the force acts. The latter factor is called *the acceleration.* In the case of the atmosphere the force is that of gravity acting in a direction normal to the earth's surface; and the acceleration of this force, usually denoted by g, at the sea-level on the parallel of 45°, if the unit of time is one second, is 9.806^m or 32.17 feet. The atmosphere being subject to the action of the force of gravity, it presses upon the earth's surface at any given place and upon itself at any level above the surface very nearly in proportion to the mass of air above. Since the force of gravity is not precisely the same at all latitudes and altitudes, being about $\frac{1}{192}$ part greater at the poles than at the equator, and diminishing a little with increase of altitude, the same mass of air does not press with quite equal force at all latitudes and altitudes.

If the surface upon which the atmosphere presses is that of a liquid, and any part of it is entirely relieved of this pressure, as in the case of a barometer tube or a suction pump, the liquid rises until the pressure of the vertical column is exactly equal to that of a column of atmosphere of the same base extending up to its top. Since the volumes of different liquids are inversely

as the densities, the greater the density the less in proportion is the volume, and consequently the height to which the liquid has to ascend, to counterpoise the pressure of the atmosphere.

10. For any given standard temperature, as that of melting ice, the height of the liquid column, either of mercury or of water, becomes a measure of the atmospheric pressure, provided this column and the atmosphere are both acted upon by the same intensity of the force of gravity, that is, force per unit mass. This latter provision becomes necessary as well as that of equality of temperature, since the pressure is measured by the product of the mass into the acceleration, and the latter, being as the intensity of force, varies slightly with a change of both latitude and altitude. A true measure must be independent of both temperature and locality.

Since the density of mercury at the standard temperature of melting ice is 13.596 times greater than that of water at the standard temperature of 4° C., which is that of its maximum density, the column of water has to rise 13.596 times higher than that of mercury, both being taken at their standard temperatures, in order to counterpoise the pressure of the atmosphere. Where the atmospheric pressure, as it usually is, is indicated by the height of the mercurial column as observed in the barometer tube, it is called *the barometric pressure.*

When the height of the mercurial column is observed under different temperatures and subject to different forces of gravity, there must be a reduction, for the reason just given, to some standard temperature and force. The assumed standard temperature is that of melting ice or zero of the Centigrade scale, and the reduction to this standard is called *the reduction to freezing.*

Since also the pressure of the same mass of mercury is different on different parallels of latitude, and at different altitudes, as has been stated, the observations of the barometer must be reduced to the height at which the mercurial column would stand if placed on the parallel of 45° and at sea-level. This is called *the reduction to standard gravity.*

The average barometric pressure for all seasons and for all

parts of the earth's surface is found from observation to be about 760 millimeters (29.92 inches). This is therefore assumed as the pressure of an atmosphere, and in cases of very great pressures is assumed as the unit of pressure, the pressure being said to be equal to a certain number of atmospheres. It is also the standard pressure, or value of P_0 in §4, in giving the absolute and relative densities of gases.

The reductions of the observed heights of the mercurial column at sea-level, which may be assumed to be sensibly the reduction of 760^{mm}, to the parallel of 45°, is given for each degree of latitude in Table I of the Appendix. For the parallels from 0° to 45° the reductions are negative and for the rest positive, as indicated by the signs placed over the first and last columns of the table.

At the poles the pressure of the mercurial column is greater than on the parallel of 45°, and hence the height at which it must stand in order to counterpoise the pressure of the atmosphere is less than it would be if subject to the standard force of gravity of the parallel of 45°, and so it must be increased in order to reduce it to this standard. At the equator the reverse is true, and hence the reduction there is negative. Between the parallels of 45° and the poles, therefore, the reductions are positive, but between this parallel and the equator, negative.

For pressures at different altitudes above the earth's surface where they are less than 760^{mm} the reductions given in the table must be decreased in proportion. For instance, on the top of Pike's Peak where the barometric pressure is only about three fifths of that at the sea-level, the proper reduction would be only about three fifths of that given in the table.

The forces of gravity are very nearly inversely as the squares of the distances from the earth's surface. Hence at any given altitude above the earth's surface the pressure of the measuring column is less than it would be at the earth's surface. There must be, therefore, a negative reduction applied for the same reason as there is at the equator, where the pressure of the mercurial column is less than it is on the parallel of 45°. This reduction in millimeters is given by the expression $-0.000003\, hP$,

in which h is the altitude of the station in meters and P the observed barometric pressure in millimeters. The exact numerical coefficient by the law above is 0.00000314, but this is diminished a little for the effect of the increase of gravity at the top by the attraction of the mountain mass. This, however, is found to have scarcely any sensible influence.[1] In English inches the expression above is $-0.0000232\,hP$, in which h must be expressed in feet and P in inches. The reductions for any given altitude are very readily obtained from these simple expressions.

11. Since force or pressure is measured by the product of the mass into the acceleration, pressure is proportional to, and becomes a measure of, mass where the acceleration is the same for both the measuring and the measured mass, as in the case of ordinary weighing, by counterpoising the one against the other, both having the same position upon the earth with regard to latitude and altitude. Two masses which have the same pressures, or, in other words, exactly counterpoise each other where they are suspended at equal distances from the fulcrum on the beam of the balance, are said to have the same weight. The weight of a body, therefore, is proportional to the mass, and not to its pressure under the varying positions which it may have with regard to the earth and other attracting bodies, and is therefore a measure of mass, and not of pressure.

Any mass which weighs as much as a cubic centimeter of pure water at the standard temperature of 4° C., which is that of its maximum density, is called a *gram*. But if this is used as a unit of pressure instead of one of mass, as it sometimes is, this unit must be understood to be the pressure of a gram subject to the standard force of gravity as already defined, since the pressure of the same mass differs in different localities. The pressure of the atmosphere in grams, as thus defined, on a square centimeter of the earth's surface is equal to the height in centimeters of a column of pure water of the temperature of 4° C. and subject to the standard force of gravity, which would exactly counterpoise it, as in the case of the mercurial column in a barometer; for the pressure would be that of a column of

water of that height and having a base of one square centimeter, and consequently the pressure in grams would be equal to the number of centimeters in the height of the column. Or, since the density of mercury at 0° C. is 13.596 times greater than that of water at the temperature of maximum density, it is equal to the height in centimeters of the mercurial column in the barometer when reduced to freezing and to standard gravity multiplied into 13.596. Hence the pressure in grams of a standard atmosphere of a barometric pressure of 760^{mm}, or 76^{cm}, upon each square centimeter of the earth's surface, is $76 \times 13.596 = 1033.3$. This upon a square meter is 10333 kilograms.

As a centimeter is equal to 0.3937 of an inch, and a gram is equal to 0.00220462 of a pound avoirdupois, the pressure of such an atmosphere in pounds avoirdupois upon a square inch is 1033.3 multiplied into 0.00220462 and divided by 0.3937^2, which gives 14.696.

12. The density of air relative to that of pure water of the standard temperature of 4° C. is called its *absolute density*. But as air is very elastic, and its elasticity depends upon both the pressure to which it is subject and its temperature, the density is a very indefinite thing unless it is given for some standard pressure and temperature. The density of pure and dry air under the standard barometric pressure of 760^{mm}, and standard temperature of 0° C., deduced from Regnault's experiments, is 0.00129278. If the density of such an atmosphere were the same at all altitudes as at the earth's surface, it would of course have a definite limit, and the height of such an atmosphere in centimeters, neglecting the very slight diminution of gravity with increase of altitude, would be 1033.3^{cm} divided by 0.00129278, or 7993 meters; for 1033.3^{cm} is the height of a column of pure water at the temperature of 4° C., which, being subject to the action of the force of standard gravity, counterpoises a column of air of the same base and of barometric pressure of 760^{mm} which extends to the top of the atmosphere; and the height of a column of air of uniform density must be to that of the column of water of the same pressure inversely as the

densities, that of water here being unity. This is called *the height of a homogeneous atmosphere.*

If the mass of the atmosphere were increased or decreased in any given ratio, the pressure and density at the earth's surface would be increased or decreased in precisely the same ratio, and consequently the height of the corresponding homogeneous atmosphere would in all cases be the same. This may also be inferred from what has been stated in § 8. For any other temperatures above 0° C. the corresponding heights of the homogeneous atmospheres would be $\frac{1}{273}$ part greater for each degree Centigrade of increase of temperature, and hence for different temperatures they would be as the absolute temperatures. With the usual amount of carbonic acid gas in the air the height of a homogeneous atmosphere is a very little less than in the case of a pure atmosphere of the same mass.

In hypothetical atmospheres of the different kinds of gases it is evident that the heights of the corresponding homogeneous atmospheres, whatever their masses, are proportionally greater for gases of greater than for those of less elastic force; for the effect of an increase of elastic force is the same whether this arises from an original greater velocity imparted to the molecules, or from an increase of this velocity and of elasticity by means of an increase of temperature. As in a static atmosphere the elastic force and pressure are equal and the densities for different kinds of atmospheres of the same mass are inversely as the elastic forces, it follows that the heights of homogeneous atmospheres of different elastic forces are inversely as their relative densities. Hence the height of a homogeneous atmosphere of pure and dry air being 7993 meters, that of a hydrogen atmosphere with a relative density of only 0.0692 would be $7993/0.0692 = 115,506$ meters.

13. The pressure of the atmosphere at any given altitude can be computed from a well-known formula where the temperature at different altitudes is known. This, for the average conditions of the atmosphere for all places and at all altitudes, decreases about 0.4° C. for each 100 meters of increase of altitude. Assuming that it decreases at this rate, and that

the temperature at the earth's surface is 20°, the temperatures at the different altitudes given in the second column of the following table will be those of the first column, and the barometric pressures at the several altitudes as given in the last column:

Temperatures °C.	Altitudes in Kilometers.	Altitudes in Miles.	Pressures in Millimeters.
+20	0	0.00	760
+10	2.5	1.56	565
0	5.0	3.11	416
−10	7.5	4.66	301
20	10.0	6.21	217
30	12.5	7.77	153
40	15.0	9.32	108
50	17.5	10.87	75
60	20.0	12.42	51
80	25.0	15.53	27
100	30.0	18.63	9
120	35.0	21.74	3
−140	40.0	24.85	1

The pressures are computed from a formula based upon Boyle's law, which, we have seen, § 3, may not hold accurately up to the altitude of 40 kilometers, where the atmosphere, by this law, becomes extremely rare; but up to the altitude of 30 kilometers, and higher, the deviations from the law are extremely small. But if there is any certain evidence that the atmosphere extends to the height of 45 miles, as is usually said, and with a density sufficiently great to indicate its existence there by means of reflected light, there must be very great deviations from this law at low pressures. With a higher temperature the atmosphere would be expanded upward more, and consequently the pressures would not decrease so rapidly with increase of altitude if our assumed temperatures in the upper part of the atmosphere in the preceding table were greater, but no probable error in these assumed temperatures would affect the results much in the last column of the table.

The following table contains a few of the observed pressures

of the atmosphere at different altitudes in different parts of the earth:

Places.	Altitudes in		Pressures in Millimeters.
	Meters.	Feet.	
Level of the Ocean	0	0	760
Geneva Observatory	408	1,339	726
Summit of Vesuvius	1,200	3.937	660
Mount Washington	1,916	6,285	600
Summit of the Great St. Bernard	2,479	8,130	565
The Summit of the Faulhorn (Bravais)	2,674	8.773	555
Summit of Etna	3.320	10,893	510
Pike's Peak	4.308	14,134	451
On the Chimborazo (Humboldt and Bonpland).	6,100	20,014	360
Glaisher's highest balloon ascent	11,280	37,000	180

These pressures are annual means in some cases, deduced from a series of years of observations, but mostly they have been obtained from one, or at most only a few, observations under different conditions of temperature and pressure at the surface, and some of the altitudes were only approximately known.

Table VI, Appendix, contains the height, in meters, of a column of dry air, corresponding to a millimeter of the barometer at different temperatures and under different barometric pressures. This will be found useful in the course of this work, and is of great practical value for various purposes. For instance, in balloon ascents and in the ascent of high mountains, with the observed barometric pressure and temperature as arguments, the table indicates the amount of ascent or descent corresponding to each change of one millimeter in the barometric pressure.

AQUEOUS VAPOR OF THE ATMOSPHERE.

14. If the atmosphere had the same temperature at all altitudes, and were in a perfectly quiet state, the aqueous vapor contained in it would form an independent atmosphere around the globe of the same nature as that of oxygen or nitrogen, § 6, and it would be distributed through the other elements in ac-

cordance with Dalton's law, and so it would exist at all altitudes in the same proportions as if the dry atmosphere were not present, and the densities would be such as are given by the laws of Boyle and of Charles, just as in the case of the other gases. But on account of its less relative density than that of the atmosphere, 0.622, it would be expanded upward more, so that it would be necessary to ascend higher than in the atmosphere, in the ratio of 1 to 0.622, before getting above the half or any other proportion of it, and it would exist in greater proportions in comparison with the other constituents, oxygen and nitrogen, in the higher altitudes than at and near the earth's surface. But it so happens that the prevailing temperatures of the atmosphere are those at which aqueous vapor of considerable tension is condensed into liquid, and as the temperature decreases with increase of altitude, the vapor tension at the different altitudes is not determined, as in the case of the other constituents of the atmosphere, by the laws of Boyle and of Charles, but by the temperatures alone which exist at these altitudes, and hence by a law which is very different. It is found from experiment that to any given temperature there corresponds a certain maximum vapor tension, which is the same whether the vapor exists alone or forms a constituent of the atmosphere, and if by any means the tension is increased above this, a part of the vapor is at once condensed and the tension reduced to the maximum corresponding to the existing temperature.

15. The maximum vapor tensions corresponding to the given temperatures, which include mostly those within the range of ordinary observation, are given in Table II. They have been deduced from Regnault's experiments by the International Bureau of Weights and Measures. The table will be convenient for reference in studying the relations between the vapor tension and the temperature, and the relations between the absolute maximum tension which can exist in the air at any given altitude and temperature and that which could exist by the laws of Boyle and of Charles.

If the temperature at the earth's surface were $25°$, and the same at all altitudes, and the vapor tension 23.517^{mm},

then, in a quiet state of the atmosphere, its distribution with regard to altitude would be according to Dalton's law, or such as would take place if the vapor were an independent atmosphere, in accordance with the laws of Boyle and of Charles. But since the temperature of the atmosphere decreases usually with increase of altitude, and at such a rate that the maximum tension corresponding to it decreases more rapidly than it would by the laws of Boyle and of Charles, the maximum tension is that of Table II corresponding to the existing temperature, and is usually less than what would be given by Dalton's law. It is seen from the table that at any altitude where the temperature is at the zero of the Centigrade scale the maximum vapor tension which can exist there is only 4.569^{mm}; whatever may be the tension at and near the earth's surface, where a warmer temperature prevails. By the laws of Boyle and of Charles, if the vapor tension were 20^{mm} at the earth's surface, at the altitude of 5 kilometers (3.1 miles) it would be about 13^{mm}; but if at that altitude there was the temperature of 0° C., it is seen from the table that the maximum tension which could exist there would be only about one third of that.

As the vapor rises or is diffused upward to altitudes where the temperature is less than that which, by the table, corresponds to its tension, the vapor is condensed and the tension reduced to the maximum corresponding to the temperature as determined by the table. When the air contains as much vapor as, at the given temperature, can exist in it, it is said to be *saturated*, and the temperature of unsaturated air at which, as it cools, the vapor begins to condense is called the *dew-point*. Where the vapor tension of the atmosphere is known from hygrometrical observations of any kind, this point is readily determined from Table II, it being the temperature in the first column to which corresponds the observed tension; for at this temperature the air is saturated, and if it cools still lower, condensation takes place. For instance, if the observed vapor tension is 17.363^{mm}, then the air, whatever its initial temperature, has to cool down to 20° C. before condensation takes place, and

this is the dew-point belonging to this assumed hygrometric state of the atmosphere.

The observed vapor tension of the atmosphere in comparison with the tension of saturation at the observed temperature is called *the relative humidity.* For instance, if, in the example above, the temperature of the air were 30° C., then the tension of saturation, by the table, would be 31.51^{mm}, and hence the relative humidity equal $17.363 : 31.51 = 0.55$, or 55 per cent.

16. The amount of aqueous vapor in the atmosphere is very variable, both at different places on the earth's surface and at different altitudes, at the same time, and in the same locality at different times. The mere diffusion of aqueous vapor through the atmosphere, as that of any one gas through another, takes place slowly; so that, as the water of the ocean and of moist land surfaces is evaporated, it penetrates through the lower strata of the air very slowly, and remains mostly near the earth's surface, unless there is an ascending current to carry it upward, and when the air is calm it may be nearly or quite saturated at the earth's surface, while it is comparatively dry at only small altitudes above it. On account also of the slow rate of diffusion, and the absence, often, of rapid convective currents, the vapor generally abounds more over water surfaces and damp places, and less over the interior and dry parts of the continents. Since, also, the temperature of the air is almost always much less in the upper than the lower strata of the atmosphere, the absolute amount of vapor which can exist in the upper strata, as we have seen in § 14, is small even when the air is completely saturated, and it generally falls much below this. For the same reason the amount of vapor in high polar latitudes, where the temperature is generally very low, is small in comparison with that in the lower and much warmer latitudes. The aqueous vapor is, therefore, very irregularly distributed, in general, both with regard to the earth's surface and also altitude, and not at all as it would be as an independent atmosphere existing alone, or as it would in a quiet atmosphere in which it would have time to be equally diffused laterally to all places, and upward, so as to satisfy Dalton's law, in case the temperature at all places

and altitudes were so great that no condensation would take place.

Since the temperature of the atmosphere is very changeable, and the absolute amount of vapor which can exist in it, by the table, depends very much upon the temperature, the actual amount found in it must also be very different at different times in the same locality, so that it is very variable both with regard to space and time.

DYNAMICAL HEATING AND COOLING OF THE AIR.

17. It is well known that if air or any gas is suddenly compressed there is an increase of its temperature, and the contrary if the pressure is removed and it is allowed to suddenly expand. The former is seen in the sudden compression of air in a condensing syringe or in a pneumatic tinder-box, in which light and dry substances may be ignited from the increased temperature, and the latter, in the rarefaction of air in the receiver of an air-pump, by which it is soon so cooled that the aqueous vapor contained in it is condensed.

The action of a force F through the space s is called *work.* If the force is constant through the whole space, the product, Fs, of the force into the space is *the measure of the work.* Where the force varies, the amount of work can only be obtained by integrating the products of each element of the spaces into the force corresponding to it. Work usually consists in moving bodies in opposition to the force of gravity, or in overcoming inertia in the case of the acceleration of the motions of bodies, or in overcoming frictional or other resistances to motion, by the application of some force.

If a body is raised vertically through the space s by the application of any kind of force, a certain amount of work has been done to place it in this position. The force which has been overcome is the force of gravity, equal gm, where g represents the acceleration of the force of gravity and m the mass of the body. The force applied is equal to that overcome, and so the amount of work done is gsm.

18. The capacity for doing work is called energy. The elevated body now, acted upon by the force of gravity, if it has space to move through, has the power of overcoming inertia, or resistance of any kind to its motion. It consequently has energy, and this is usually called *potential energy*, sometimes *energy of position*, and is exactly equal to the amount of energy spent, of whatever kind, in raising it to its position. If the body in falling is not obstructed in its motion by some resisting medium or frictional resistance, this potential energy is all spent in overcoming the inertia of the body in producing accelerated motion. Letting v represent the velocity of motion corresponding to the space s passed over, we have by the well-known relations between the space and velocity in the case of falling bodies, $gs = \frac{1}{2}v^2$, and so the amount of energy spent in producing this velocity or the momentum mv, is $gsm = \frac{1}{2}v^2m$. But this body in motion now has the capacity of overcoming other forces or resistances, and consequently has energy, called *kinetic energy*, formerly *vis viva* or living force, and the measure of this energy is equal to that of the work gsm spent in producing it, and consequently equal to $\frac{1}{2}v^2m$. The kinetic energy is therefore proportional to the square of the velocity.

It is well known that heat is produced by the concussion of inelastic bodies. If the body, of mass m, after having acquired in any way the velocity v, should be brought to rest by contact with another body, as in the case of completely inelastic bodies, it would then have lost all of its kinetic energy. This kind of energy is thus transformed into heat, which is a well-known agent in doing many kinds of work, and is called *thermal energy*. If the velocity v is produced by potential energy, then this is directly transformed into kinetic energy, and thus indirectly into thermal energy.

If a falling body or a body descending in any way from a higher to a lower level, is retarded by the action of a resisting medium or by friction, then the potential energy is spent partially in giving momentum or kinetic energy to the body and to certain parts of the fluid, and partly in causing heat. But as soon as the body is brought to rest, and likewise all parts of

the fluid in the case of a resisting medium, then all the kinetic energy generated at first, is likewise transformed into heat, and the whole amount is proportional to the amount of potential energy of the body spent in descending to a lower level. This heat is contained partly in the body acted upon, and partly in the resisting medium or body offering the frictional resistance.

19. We have seen that a given amount of potential energy can be transformed into a corresponding amount of kinetic energy. Conversely, this kinetic energy can be transformed back again into potential energy, as, for instance, where the whole momentum acquired is used up in causing the body to ascend against the force of gravity, in which case the body ascends exactly to the height from which it had fallen in generating the kinetic energy, and so the same amount of potential energy is reproduced which was spent in producing the kinetic energy. But there must not be any resisting medium or frictional resistance affecting either the descending or ascending motion of the body. In like manner the heat generated by a given amount of kinetic energy can be used to reproduce this energy, and if it could be applied without loss it would give rise to exactly the same amount of kinetic energy as was spent in producing it. Hence each of these three kinds of energy can be transformed, either directly or indirectly, into each of the others; and the amount of energy in each case, if it could be applied in doing work without loss, would do the same amount of work. The same is true with regard to other kinds of energy which might be named. In fact, work is simply the creation of energy of some kind at the expense of another, or the transformation of one kind of energy into another. In this way, at the expense of either heat or kinetic energy, the work of raising a body against the force of gravity is performed, or, in other words, potential energy is produced. The same, in doing the work of overcoming frictional resistances to motion, generates as much heat as is spent, though it may be so scattered as not to be perceptible.

In the compression of air a certain amount of work is done at the expense of some kind of energy. This is transformed

for the most part directly into heat, the amount of kinetic energy in the motion of the air in being compressed, and which is transformed into heat energy as soon as all motion ceases, being very small. If heat energy were the agent used in the compression, and it could be applied without loss, the heat of the compressed air would be increased by the amount of heat expended in the compression. The whole amount of energy, therefore, remains the same, and is not diminished in doing work. This principle is called *the conservation of energy.*

20. If the whole potential energy of a raised body, subject to the action of the force of gravity, is spent in falling, by means of some mechanical contrivance, in agitating a given amount of water, and so heating it by means of friction, the amount of heat generated is the equivalent of the potential energy spent, and so of the work required to raise the body. In this way we can compare the work with the heat produced by its equivalent potential energy, and hence get an equivalent expression of one in terms of the other, whatever the units used in the measure of each. The raising of one kilogram, subject to the force of standard gravity, through one meter, called a kilogram-meter, is taken as *the unit of work.* The *unit of heat* is the amount of heat required to raise the temperature of a unit mass of pure water from 0° to 1° C. The unit of water which is usually assumed in the unit of work is the kilogram, and so the heat unit in this case is the amount of heat required to raise the temperature of a kilogram of pure water from 0° to 1° C. According to the most recent experiments,[4] the height to which a unit of heat is capable of raising one kilogram, subject to the force of standard gravity, may be put at 430 meters, the equivalent of which would be the raising of 430 kilograms through one meter. Hence the capacity of such a heat unit for doing work is 430 kilogram-meters. This is called *the mechanical equivalent of heat.* The amount of heat, therefore, required to do one unit of work, called *the equivalent of work,* is the reciprocal of this number, or $\frac{1}{430}$ of a unit of heat.

21. If heat is applied to any given portion of air which is

free to expand, a part of this heat is consumed in the work of expansion and the balance only goes toward heating the air. Taking a simple case, if we have a cubic meter of pure and dry air at the standard temperature of 0° C., and under the standard pressure of one atmosphere, and free to expand in one direction only, the heat which raises this cubic meter of air through one degree of temperature, we have seen in § 4, expands it in this one direction $\frac{1}{273}$ part of one meter. The resistance to the expansive force in the expansion is the pressure of a standard atmosphere on a square meter of surface, which is 10,333 kilograms (§ 11). The space through which this resistance is overcome being $\frac{1}{273}$ of a meter, the amount of work done in the expansion in kilogram-meters is $10333 \times \frac{1}{273} = 37.85$.

If the cubic meter of air should expand equally in each of the directions normal to its six surfaces, then the space through which each surface of the cube would move, for the same expansion of volume, would be only $\frac{1}{6}$ as much, but the amount of surface moved and force overcome would be six times as much, and so the amount of work done would be precisely the same. Moreover, whatever may be the figure of the volume of air equal to that of one cubic meter, and the amount of surface, and the amount of displacement of this surface in the various directions perpendicular to the surface, the same amount of work is done in the same expansion of volume; for the work done is proportional to the sum of all the elements of surface multiplied into the space through which they move in the expansion, and the pressure or resisting force being the same on all sides, the amount of work done is the same in all cases for the same expansion of volume.

22. The same masses of different bodies at the same temperature require different amounts of heat to raise their temperatures through one degree. By the definition of a heat unit, § 20, it requires one unit of heat to raise the temperature of a kilogram of pure water from 0° to 1° C.; but it is found from experiment that to raise the temperature of a kilogram of pure and dry air, under constant pressure, from 0° to 1° C., only

0.2375 of a unit of heat is required. This is called *the specific heat of air.*

According to the results obtained by the International Bureau of Weights and Measures from the experiments of Regnault, a cubic meter of pure dry air of standard pressure and temperature weighs 1.29278 kilograms. Multiplying this into the specific heat of air, we get $1.29278 \times 0.2375 = 0.30703$ of a heat unit as the amount required to increase the temperature of the air by one degree, and so to expand the cubic meter of air $\frac{1}{273}$ part of its volume, since it expands so much for each degree of increase of temperature. But we have just seen that in this expansion the amount of work done is 37.85 kilogram-meters. Dividing this therefore by 0.30703, we get 123.28 kilogram-meters as the amount of work done by a unit of heat in expanding the cubic meter of air the $\frac{1}{273}$ part of its volume.

If the cubic meter of air, of whatever figure, were subject to any other than the standard barometric pressure of 760^{mm}, then the work corresponding to any given amount of expansion of the volume would be in proportion to the pressure, and consequently to the density, where the temperature is constant. But the amount of heat required to raise the temperature of a given volume of air through one degree is likewise in proportion to the density. Hence the same amount of heat is required to be applied to the cubic meter of air to do a certain amount of work of expansion, whatever the pressure and density may be.

Where the cubic meter of air has temperatures differing from that of 0°, the amounts of expansion, and consequently of work done, by the heat which raises its temperature by a given quantity, are inversely as the absolute temperatures, being a constant part of the volume when reduced to the temperature of 0°. But the amounts of heating by a given quantity of heat are *directly* as the absolute temperatures, being inversely as the densities. The same quantity of heat, therefore, applied to the cubic meter of air at all temperatures does the same work of expansion.

Again, if a unit or any other quantity of heat is applied to any other volume than a cubic meter under the same circum-

stances, there is the same work of expansion done, for the greater the volume, of whatever figure, the greater the surface, but the less in proportion is the heating and the space through which this surface is moved, by the application of the same amount of heat.

From what has been shown, it now follows that the amount of work of expansion done by a given quantity of heat is the same whatever the pressure, temperature, and volume may be to which it is applied, and in all cases the amount of work done by a unit of heat is 123.28 kilogram-meters. But the whole work equivalent of this heat unit is 430 kilogram-meters, and therefore the part of the heat used in doing the work of expansion is to the whole as 123.28 to 430, or as 0.2867 to unity. Of a unit of heat, therefore, applied to any given volume of air, 0.2867 of a unit is used in doing work, and the balance, 0.7133 of a unit, goes toward heating the air.

23. Where the air is not allowed to expand there is no work done at the expense of the heat supplied, and the whole goes toward heating the air. The increase of temperature, therefore, given to air by a certain amount of heat communicated to it where it is not allowed to expand, is to that in the case of expansion under constant pressure in the ratio of unity to 0.7133, or as 1.402 to unity. Conversely, the heat required to raise the temperature of a given quantity of air through one degree under constant volume is to that required in the case of expansion under constant pressure as unity to 1.402. The specific heat of air, therefore, under the latter conditions being 0.2375, in the case of constant volume it is $0.2375/1.402 = 0.1694$.

We have seen that of the heat communicated to air under constant pressure which raises its temperature one degree, the part 0.2867 is used in doing the work of expansion. Consequently, if no heat is communicated, and the same work of expansion is done in the air's coming under different pressures, this work is done at the expense of the heat already in the air, and hence it is cooled $0^\circ.2867$ by this expansion.

24. In order to obtain the heating effect of compression, or

the cooling effect of expansion, we must compute the amount of work spent in compression, or of work done in expansion, and then the equivalent of this work in heat units is equal to the number of kilogram-meters of work divided by 430. Each one of these heat units produced by the expenditure of work in compression would raise the temperature of a kilogram of water one degree, but as the specific heat of air under constant volume, in which case no heat is spent in doing work, is 0.1694, the temperature of a kilogram of air would be raised in the ratio of unity to 0.1694, or to 5°.9 by each unit of heat produced by the work of compression. Each unit of work, therefore, would heat a kilogram of air the $\frac{1}{430}$ of 5°.9, or 0°.0137. On the contrary, for each unit of work done in expansion at the expense of the heat in a kilogram of air, it is cooled by the same amount.

In the compression or expansion of confined air the work expended in the one case, or done in the other, is not proportional to the amount of contraction or expansion of the volume, since the force, or resistance overcome, is constantly changing, becoming continually greater in case of compression, and less in that of expansion, so that the amount of work can be computed accurately only by means of a logarithmic formula. Under the special case, however, where heat is communicated to, or taken away from, air under constant pressure, the force is constant, and the work done in the expansion, or spent in compression, is proportional to the change of volume.

25. We now come to the subject of the cooling or heating of air in ascending or descending currents. The amount of expansion of air under constant pressure corresponding to an increase of temperature of 1° is $\frac{1}{273}$ of the volume at the temperature of 0° C. Hence wehn the volume of air in coming under less pressure expands so rapidly that there is no time for it to gain or lose heat by radiation or conduction, it loses 0°.2867 of temperature for each expansion of $\frac{1}{273}$ of its volume at the temperature of 0° C., and consequently this expansion must be increased in the ratio of unity to 0.2867, or to $\frac{1}{78.27}$ part of its volume in order to cool one degree.

It has been shown, § 11, that the height of a homogeneous atmosphere, at the temperature of 0° C., reckoned from the earth's surface or any other level, is 7993 meters. If, therefore, air of this temperature ascends one meter, the pressure is diminished, and the volume increased, the $\frac{1}{7993}$ part. Hence, dividing $\frac{1}{78.27}$ by $\frac{1}{7993}$, we get 102.1 meters for the height to which the air must ascend to cool one degree. This gives a rate of cooling of air in its ascent of 0°.98 very nearly for each 100 meters of ascent.

For other temperatures than that of 0° C., the masses of air left below in ascending one meter, and consequently the diminutions of pressure, are inversely as the absolute temperatures, and consequently the amounts of expansion of volume at those temperatures are in the same proportion. But the volumes for different temperatures are *directly* as the absolute temperatures. The air therefore in ascending one meter, whatever its temperature, is expanded the $\frac{1}{7993}$ part of its volume at the temperature of 0° C., and hence is cooled by the same amount as in the case of air of the temperature of 0° C. For all temperatures of the air, therefore, it has to ascend through 102.1 meters to cool 1°, and hence cools 0°.98 for each 100 meters of ascent.

In the case of air descending vertically we have compression instead of expansion, and consequently a heating instead of a cooling, and at the same rate. Since the cooling in the one case depends upon expansion, and the heating in the other upon compression, it is not necessary that the ascent or descent should be vertical, but only that the air should come under different pressures, and consequently arrive at different levels, and the rate of cooling or heating is 0°.98 for each 100 meters of change of level, whatever be the directions of ascent or descent.

26. Where the ascending air is saturated with aqueous vapor, the cooling arising from the expansion continually diminishes the capacity of the air for moisture, for by Table II the lower the temperature the less the amount of vapor which can exist in the air. Hence, as the saturated air ascends and

cools, the aqueous vapor is condensed, and the latent heat gradually given out, as condensation takes place, raises the temperature of the air above what it would otherwise be, and thus causes the rate of cooling of the ascending saturated air to be much less than in the case of dry or unsaturated air.

The number of heat units required to evaporate one unit by weight of water is $606.5-0.695\tau$, in which τ is the temperature above that of melting ice. Hence it decreases with increase of temperature, and in the case of boiling water, in which $\tau = 100°$, it becomes 536. This means that at the boiling temperature the heat which is used in the transformation of water into vapor would heat 536 units of water at the zero temperature by one degree. Condensation being the reverse of evaporation, where the vapor in the air is gradually condensed in ascending, each unit by weight of vapor condensed gives out a certain amount of heat, varying a little by the preceding expression with the temperature.

From an inspection of Table II it is seen that the colder the air the less is its capacity for moisture, and also the less is this decreased by a given decrease of temperature. For instance, at the temperature of 30° the vapor tension of saturation by the table is 23.517^{mm}, and in the ascent of the air and cooling down to 25° the tension is diminished to 17.363^{mm} by condensation. But when saturated air has a temperature of only 5° the vapor tension is 6.507^{mm}; and in ascending and cooling down to 0° the tension is reduced by condensation to 4.569^{mm}, and so is decreased by 1.938^{mm} only, and consequently there is in this case a much less amount of vapor condensed than in the other for the same decrease of temperature. Since, therefore, the rate of decrease of temperature with increase of altitude in the case of no condensation is the same at all altitudes and for all temperatures, and the amount of condensation and of latent heat given out for the same amount of cooling is much less for the lower than for the higher temperatures, in the ascent of saturated air at very low temperatures, such as generally prevail in the winter season, and at all seasons in high altitudes, the rate of cooling with increase of altitude in ascending air is

much greater than it is in the summer season, since in the former the latent heat given out is much less and the rate becomes more nearly that of ascending dry or unsaturated air, namely, about one degree for each 100 meters.

The rate of cooling in ascending saturated air also depends upon the density of the air and consequently upon the altitude. At high altitudes the rate of condensation in ascending saturated air, for the same temperatures, is the same as at low altitudes for equal volumes, and consequently the rate with which heat is given out as the air ascends is the same in both cases; but at high altitudes, where the density is less, the same amount of latent heat given out in condensation gives rise to a greater increase of temperature, the heating being inversely as the density; and hence at these altitudes, at the same temperatures, the rate of decrease of temperature in ascending air is less, since the same volumes of air there are heated more by the same amount of condensation. In the higher altitudes, however, the temperature is usually much less, and on this account the rate of decrease of temperature above is not nearly so much diminished as it would be if the temperature there were the same as at lower altitudes.

After the saturated air has ascended to an altitude where it has cooled down to the temperature of 0°, the vapor is then condensed into snow, and by this an additional amount of latent heat is given out equal to that of liquefaction, which is 79.25 heat units for each unit of weight of liquid, and so is about one seventh of that given out in the condensation of the vapor into rain. This diminishes a little the rate of cooling above the plane of incipient freezing.

Where there are rain-drops and fine cloud-particles carried up above this plane, as there always are more or less in ascending currents, the rate of cooling is still further diminished by the latent heat given out in freezing, until all are completely frozen, and this rate may even be sensibly suspended for some little distance above this plane, where the water particles carried up are very fine, and so thoroughly diffused through the air that the heat given out is almost instantly communicated to all parts,

for until the drops and fine particles are completely frozen they must remain at the temperature of 0°.

In descending currents of air, even if it is saturated at the start, the rate with which it becomes heated with decrease of altitude is in all cases the same as that of dry air—about one degree for each 100 meters; for if it is saturated it at once becomes unsaturated as it descends and becomes warmer, and consequently there is no condensation, and the rate is in no way affected by latent heat becoming sensible, as in the case of ascending saturated air. Where the saturated air is clouded, the clouds consisting of very small droplets of condensed vapor, these, after the air in descending becomes unsaturated, are gradually evaporated, and the heat of evaporation taken from the air causes the rate of increase of temperature in descending to be a little less than that of unsaturated and clear air, until the cloud particles are all evaporated and the air becomes clear.

27. The theoretical formula for computing the rates of decrease of temperature of saturated ascending air with increase of altitude, for the different conditions with regard to temperature and altitude, is too complex to be investigated here, but these rates are given in Table III of the Appendix corresponding to the temperatures at the heads of the columns, and the pressures in the first, and the approximate altitudes in the last, column of the table.[a]

It has been explained in § 26 that the rate of decrease must be greater for the same pressures and altitudes in the case of low than in those of high temperatures, and also, that for the same temperatures the rates must be greater near the earth's surface than at considerable altitudes. It is seen from an inspection of the computed rates in the table that this is the case; for at the earth's surface for a temperature of −10° this rate is twice as great as it is for a temperature of 30°, and the rates also for the same temperature, are much less above than at the earth's surface.

From Table III we readily compute the temperature of rapidly ascending saturated air at any given altitude when it is

known at the earth's surface or any other lower level. For instance, suppose the temperature at the earth's surface is 30°, and the temperature of the ascending current is required at the altitude of 4000 meters (2.5 miles nearly). It is seen from the table that the rate of decrease at the earth's surface at that temperature is 0°.37 per 100 meters, and that this is very nearly the rate at all altitudes in its ascent, since the air as it ascends gradually becomes cooler. It can therefore be assumed to be the same in getting approximate values of the rate of decrease of temperature at the altitude of 4000 meters. The temperature, therefore, at the altitude of 4000 meters is decreased approximately $0°.37 \times 40 = 14°.8$, and hence the approximate temperature at that altitude is $30° - 14°.8 = 15°.2$. Corresponding to this temperature and the altitude 4000 meters the table gives a rate of decrease of 0°.39. Taking the average of this rate and that at the earth's surface, which is 0°.38, we get a more correct decrease of temperature at the height of 4000 meters, $0.38 \times 40 = 15°.2$, which being deducted from the surface temperature, 30°, we get 14°.8 for the temperature at that altitude.

Where the air at the earth's surface or any of the lower strata of the atmosphere is not saturated with aqueous vapor, but only becomes so after having ascended to a certain altitude, the rate of decrease before arriving at this altitude is 0°.98 per 100 meters of ascent. This altitude is where the air in ascending becomes cooled down to the dew-point which it has after having reached that altitude, which is lower than the dew-point of the lower level from which it has ascended, since the vapor tension of the lower level is decreased from expansion, and this in the ratio of the decrease of air-pressure. The vapor tension at the lower level being given, the dew-point of the level of incipient condensation is that temperature of Table II corresponding to this tension diminished in the ratio of the pressures of the two levels. The ratio between these pressures, and also the amount of cooling of the air in ascending, depend upon the height of ascent, and this for the plane of incipient condensation is determined by the condition that

the air in ascending must be cooled, at the rate of $0°.98$ for each 100 meters, down to the *new* dew-point. The altitude which satisfies this condition, where the temperature of the air τ, and the depression of the dew-point $\tau - d$ are given, cannot be determined directly, but only by approximations. These altitudes, so determined, corresponding to the air temperatures at the heads of the columns and the depressions of the dew-point in the first column, are given in Table IV.

It is seen that these altitudes are nearly independent of the air temperatures, being only a little less for freezing temperatures than for high summer temperatures. They are also entirely independent of the air-pressure, so that the lower level may be either sea-level or of any high plateau, or high level in the open air above the earth's surface.

It is also seen that if the depression of the dew-point, $\tau - d$, is multiplied into 125, we have approximately the altitudes in the table, especially for the lower altitudes. A convenient rule, therefore, for determining the altitude approximately in meters, is to add one fourth part to the depression of the dew-point in Centigrade degrees and multiply by 100. For instance, if the difference between the air temperature and the dew-point were 10°, then the air would have to ascend about 1250 meters before it would become saturated, and condensation would take place, and only after this altitude would be reached would the rate of decrease of temperature be as given in Table III, and the amount of decrease of temperature through the remaining difference of altitude be obtained as in the preceding example.

It must be understood that the rate of cooling which has been given for ascending dry or unsaturated air, as likewise the rates given in Table III, are applicable in the case only in which the ascent of air is so rapid that it does not have time to be sensibly affected in other ways than by expansion and the latent heat of condensation.

The observed rate of decrease of temperature with increase of elevation in the lower part of the atmosphere has been found to be $0°.60$ for each 100 meters on the average for all localities.

and seasons of the year, but this rate varies considerably in different places, and is thought to be in general a little greater in the lower than in the higher latitudes, and it is also considerably greater in summer than in winter. In the latter season, and especially at night in the lower strata of the atmosphere in the interior of the continents, it not only becomes very small, but may even be reversed, so that there is an increase instead of a decrease with increase of altitude.

The following table, given by Dr. Hann, contains the rates of decrease of temperature in summer as deduced from Glaisher's observations during balloon ascensions:

DIMINUTION OF TEMPERATURE PER 100 METERS IN CELSIUS DEGREES.

Weather.	Altitudes in thousands of English feet.							
	0–1	1–2	2–3	3–4	4–5	5–10	10–15	15–20
Clear.	0.98	0.71	0.55	0.55	0.55	0.46	0.39	0.30
Cloudy.	0.86	0.73	0.73	0.56	0.55	0.45	0.40	0.25

STABLE AND UNSTABLE EQUILIBRIUM.

28. If, when a portion of the atmosphere in a state of static equilibrium receives, from any slight temporary cause of disturbance, an upward or a downward motion, the changed conditions arising from such a movement tend to bring it back again to its original position, it is said to be in a state of *stable* equilibrium, since, immediately after such disturbance, the part disturbed is brought back and soon settles in its original position. But if, after any such disturbance, the changed condition tends to continue the motion and to increase the velocity of the air disturbed in the direction in which it was started, either upward or downward, the atmosphere is then said to be in a state of *unstable* equilibrium, since although in static equilibrium as long as it is not disturbed, yet the slightest disturbance introduces an initial change in the conditions, which tends to con-

tinue the motion, until by a complete inversion of the strata of the atmosphere, or from some other cause, the conditions are so changed that the state of stable equilibrium is brought about.

It is well known that if any portion of a fluid, or a solid immersed in a fluid, has a *less* density than the surrounding part of the fluid at the same level, it tends to rise up, and if not hindered, continues to rise until it comes to the top; but if, on the other hand, it has a *greater* density, it sinks down to the bottom. So if, when any portion of the atmosphere receives an upward motion, its density becomes greater, or a downward motion, less, than that of the surrounding air at the same level, it at once comes back to its former position after the temporary cause of disturbance ceases; but if, in rising up, the density becomes less, or in sinking down, greater, than that of the surrounding parts of the same level, the tendency is to continue on in the direction in which it is started. In the former case it is in a state of stable, and in the latter of unstable, equilibrium. But since a difference of density in these cases depends mostly, if not entirely, upon a difference of temperature of the ascending or descending air and that of the surrounding part of the atmosphere at the same level, the state of stable or unstable equilibrium depends very much upon the relation between the rate of decrease of temperature with increase of altitude in the surrounding undisturbed part of the atmosphere and that of the ascending or descending current.

29. We have seen that the temperature of rapidly ascending unsaturated air decreases very nearly one degree for each 100 meters of ascent. If, therefore, the temperature of the surrounding undisturbed part of the atmosphere decreases with increase of altitude at a rate less than this, then as the air, started by some slight temporary impulse, ascends, its temperature becomes less, and consequently its density greater, than that of the surrounding undisturbed air at the same level; and hence the motion is soon arrested and reversed, and the air after a few oscillations is brought by friction to a state of rest in its original position. For instance, let us suppose that the arrangement of temperature with regard to altitude, of a dry or unsatu-

rated and undisturbed atmosphere to be represented in the third column of the following table, in which the first column contains the altitudes, and the second the temperatures of the ascending current of air at the altitudes in the first column:

Meters.			
3000	0°	6°	− 6°
2500	5	10	0
2000	10	14	6
1500	15	18	12
1000	20	22	18
500	25	26	24
000	30	30	30

The temperature of the ascending column is here supposed to decrease exactly 1° for each 100 meters. It is readily seen that with the arrangement of temperature in the third column a slight upward motion of air at any place would cause its temperature at all altitudes to be a little less and consequently its density a little greater than that of the surrounding undisturbed air at the same levels, and this would continue to increase as the air ascends, until the temperatures would finally be as in the second column of the preceding table, if the ascent were to continue. But this increased pressure of the whole column having an ascending motion would soon bring it back. It would then descend a little lower than its original undisturbed position, and in so doing it is readily seen that the temperature then would be greater and the density less than that of the surrounding undisturbed air, and the further it descended below this position, the more so, and hence the downward motion would soon be stopped and reversed, and after a few oscillations it would be brought to rest by friction in its first position. And this would be the case whatever the direction, upward or downward, of the first impulse. Hence with this vertical distribution of the temperature, the atmosphere is in the state of stable equilibrium.

But if we suppose the temperature to decrease with increase of altitude at the rate of 1°.2 for each 100 meters, instead of 0°.8 as in the preceding case, we shall have the vertical dis-

tribution of temperature as given in the fourth column of the preceding table. It is readily seen that now a slight upward motion would cause all parts receiving such a motion to have a little greater temperature and less density than those of the surrounding undisturbed parts, and the greater the disturbance the less the density relative to that of the surrounding air at the same levels. Hence there would be no tendency to fall back again, but to continue on, and the further the displacement the greater this tendency until a regular ascending current would be established, in which the temperatures at different altitudes would be as represented in the second column of the preceding table. The difference of temperature, then, between the ascending air and the surrounding quiet air would increase one degree for each 500 meters, so that at the altitude of 1500 meters it would be 3°, at 3000 meters 6°, and so on; and consequently there would be a strong tendency to rush up to the top of the atmosphere, at least if the same vertical decrease of temperature in the undisturbed atmosphere should extend all the way up. If the initial motion were downward, the temperature would become less and the density greater than in the surrounding air at the same level, and hence, being once started, the tendency would be to continue. In this case, however, the down-rushing air could not escape so readily on account of the friction of the earth's surface in its lateral escape at the surface. Without some initial disturbance, however, either upward or downward, the atmosphere would remain at rest. With the vertical distribution of temperature, therefore, given in the last column of the table, if the air should receive a start, either upward or downward, the tendency would be to continue on, and hence with this distribution the atmosphere is in the state of unstable equilibrium.

If the vertical temperature gradient in the undisturbed atmosphere is such that the temperature decreases 1° for each 100 meters of increase of altitude, then the atmosphere is said to be in the *indifferent state of equilibrium.*

30. Again, let us imagine a portion of air confined within a balloon of very fine silk, the weight of which would be insensible

in comparison with the weight of the confined air. If the decrease of the temperature of the air with increase of altitude were 1°.2 for each 100 meters, then the air in the balloon in its ascent, its initial temperature being supposed to be the same as that of the surrounding air, would become 0°.2 warmer than the surrounding air for each 100 meters of ascent, or one degree for each 500 meters, and after having ascended 2000 meters would be 4° warmer than that of the surrounding air; so that the higher it would ascend, the stronger would be the tendency to continue on. But if the balloon were at some altitude above the earth's surface and started downward in the same condition of the air, the air contained within would become one degree colder for each 500 meters of descent, and consequently heavier than the surrounding air, and so, being once started, the tendency would be to go on. Supposing the silk to have no weight, and the initial temperature of the air within and without to be the same, there would be no tendency to move either up or down until once started by some initial impulse. This would, therefore, be a state of unstable equilibrium for unsaturated air.

If the rate of decrease of temperature in the air with increase of altitude were only 0°.8 for each 100 meters, then, it is readily seen, the air in the ascending balloon would become 0°.2 colder than the surrounding air for each 100 meters of ascent, and 0°.2 warmer for each 100 meters of descent: so that in the former case it would become heavier than the surrounding air, and soon cease to ascend, and then fall back again; and in the latter it would become lighter, and soon cease to fall, and would run up again to the position which it left. This would therefore be a state of stable equilibrium for unsaturated air.

It will be readily understood from the preceding illustrations that if an unsaturated atmosphere has a temperature which decreases with increase of altitude at a rate which is less than one degree (accurately 0°.98) for each 100 meters, it is in the stable state; but if this rate is greater than one degree for each 100 meters, it is in the *unstable* state.

31. If the ascending air is saturated with aqueous vapor, it

is seen from Table III that the rate of decrease of temperature with increase of altitude required to produce the state of unstable equilibrium is much less, and this is especially the case in the summer season, and in all seasons at great altitudes. For instance, in the example given in § 27, the average rate of decrease is $0^\circ.38$ for each 100 meters, and the rate is very nearly constant at all altitudes. If, therefore, the rate of decrease in the quiet air were only a little greater than this, the air would be in the unstable state, whereas in the case of dry or unsaturated air the rate would have to be more than twice as great. But with a temperature of -10° the rate of decrease in ascending air is about twice as great, and so the rate of decrease in the undisturbed and quiet air would have to be greater than this is, to give rise to the unstable state, but still considerably less than in the case of dry or unsaturated air.

In the case of saturated air the initial motion of disturbance must be upward, since when it is downward the air becomes warmer, and so at once unsaturated, so that the tendency to a continued downward motion can only take place, although the air at first may be completely saturated, when the rate of increase or decrease with change of altitude in the undisturbed surrounding air is greater than $0^\circ.98$ for each 100 meters.

RELATION BETWEEN CHANGES OF ALTITUDE AND DENSITY.

32. In a quiet atmosphere at the temperature of 0° C., the pressure, and consequently the density, is decreased by the $\frac{1}{7993}$ part for each meter of increase of altitude, for in an ascent through a vertical distance of one meter this part of the mass is left below; and this is true for all pressures and altitudes, if the temperature is the same, since the height of a homogeneous atmosphere is the same, whatever the mass and pressure. But we have seen that the volume of the air at all temperatures is increased, and consequently its density decreased, by the $\frac{1}{273}$ part of that at the temperature of 0° C., for each degree of increase of temperature. Dividing, therefore, $\frac{1}{7993}$ by $\frac{1}{273}$, we get $\frac{273}{7993} = 0^\circ.0348$ for the decrease of temperature for each

meter of increase of altitude required to give the same density at all altitudes.

For other and higher temperatures it would be necessary to ascend through a vertical distance of more than one meter in the ratio of the absolute temperatures, in order to diminish the density $\frac{1}{7993}$ part; and hence the decrease of temperature in this case would be 0°.0348 for a change of altitude greater in the same proportion, and the rate of decrease required would be inversely as the absolute temperature.

THE VISCOSITY OF THE AIR.

33. That property of the atmosphere, or of any gas by which one stratum cannot move with greater velocity over another without receiving resistance from it, is called *viscosity*. By the kinetic theory of gases this arises from the continual interchange of molecules between the strata. Those of the more slowly moving stratum, coming in contact with those of the contiguous stratum moving with greater velocity, tend to diminish its velocity, while those of the more rapidly moving stratum, coming in contact with those of the other, tend to accelerate its motion, so that there is a tendency in these contacts to equalize the velocities of the two, but leaving the whole amount of momentum in the two strata unchanged, when there is simply a mutual action between the two. The *measure* of viscosity is the amount of force required per square unit of surface to overcome the mutual action of two strata of unity of thickness upon each other, and having a relative velocity of unity. This mutual action between the two, usually called *friction* or *frictional resistance*, is proportional to the relative velocities of the strata, and, according to the deduction of Maxwell from the kinetic theory of gases, is independent of density and pressure where the temperature remains the same.

When the whole atmosphere from bottom to top moves in the same horizontal direction with velocities increasing with increase of altitude, the lower stratum in contact with the earth's surface is retarded by the contacts of its molecules with

the asperities of that surface, but the motions of each successive stratum above is acted upon by the more slowly moving stratum below and retarded in its motion, unless there is some constantly acting force to overcome the effect of this frictional resistance, and maintain a constant relative velocity between the two strata. If there are three successive strata, of which the relative velocity between the first and second is the same as that between the second and third, then the actions of the first and third upon the intermediate one are precisely the same, but in contrary directions.

Since an increase of temperature increases the velocities of the interchanging motions of the molecules between the strata, the viscosity of the atmosphere is increased by increase of temperature. Experiment makes it about as the three-fourths power of the absolute temperature.[*]

CHAPTER II.

THE MOTIONS OF BODIES RELATIVE TO THE EARTH'S SURFACE.

34. THE motion of a body relative to the earth's surface, called *relative motion*, whether this body is set in motion by a single impulse, or is continually acted upon by some force other than that of gravity, is very different upon the earth with a rotation upon its axis from what it would be if the earth were at rest. If the earth were at rest, and spherical, as it would have to be in this case, a body set in motion in any direction upon its surface, supposed to be perfectly smooth and without friction, would continue to move uniformly with its initial velocity in a great circle of the sphere. But where the earth has a motion of rotation upon its axis, we shall see that this is far from being so, except in the case where the initial motion is in the line of the equator. If, also, the earth were at rest, a body projected vertically upward would fall back to the point where it started, but in the case of the earth with a rotation upon its axis, it falls to the earth a little west of this point, and projectiles generally, started in any direction relative to the earth's surface, would have different relative motions in the two cases. In considering, therefore, the relative motions of bodies on the earth's surface, it is very important to take into account the influence of the earth's rotation upon such motions, and to not regard these as being the same as the absolute motion in case of no rotation.

CENTRIFUGAL FORCE.

35. In determining the influence of the earth's rotation upon relative motions upon and near its surface, the centrifugal force arising from the earth's rotation comes in as an impor-

tant consideration. It is necessary, therefore, to understand at the start, the nature and law of this force. If a free body in space has a motion in any direction and is not acted upon by any force, or affected by any kind of resistance, frictional or otherwise, it continues to move on with uniform velocity in this direction. If, therefore, a body at *a*, Fig. 1, has, under these conditions, a motion in the direction of the line *ad*, tangent to the circle *aegh*, of which *o* is the centre, and has a velocity which will carry it to *b* in a unit of time, as one second, then, at the end of this unit of time, it will have departed from

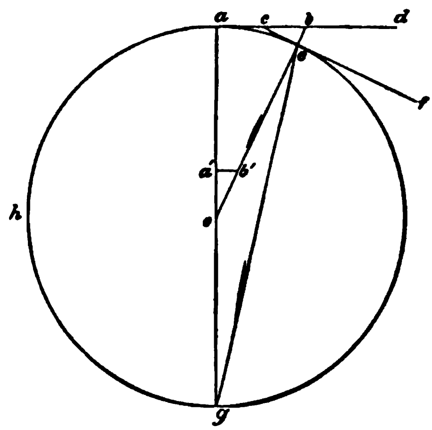

Fig. 1.

the circumference, and have increased its distance from the centre *o*, by the space *eb*. Drawing now the line *cf* tangent to the circle at *e*, it will perhaps be evident to most readers, without a rigorous demonstration, that the angle *dcf* at the intersection of the two tangents to the circle at *a* and at *e* is equal to the angle *aoe*, since at the start as the movable radius *oe* changes its direction with regard to *oa*, the tangent at *e* must change its direction with reference to that of *ad* at the same rate. In the case of very small angles between two lines forming an angle as *cd* and *cf*, the rates at which a body moving along one of them with a given velocity departs in different

cases from the other are sensibly proportional to the angles between them. The body then, in moving from *a* towards *d* with a given uniform velocity, will, at the ends of different units of time, if the unit is very small, and this is entirely arbitrary, depart from the line *cf* and from the centre of the circle with velocities which are proportional to the angle *dcf* at the ends of these units of time, and consequently proportional to the angle *aob* which is proportional to the time; for while the angle *aob* is small it is proportional to *ab*, which, since the body moves with uniform velocity, is proportional to the time. The law of departure, then, for very small arcs, is that the rate or velocity of departure of the body from the circumference of the circle and from the centre, is proportional to the time elapsed.

If a body is free and is acted upon by any constant force, as that of gravity, its motion is accelerated in proportion to the time, the increase of velocity in a unit of time being called the *acceleration*. For instance, a body near the earth's surface, acted upon by the force of gravity, acquires a velocity of about 32 feet per second at the end of the first second, 64 feet at the end of the second second, and so on; the velocity continuing to be increased in proportion to the time. Now we have seen above that in the motion of a body in the direction of a tangent to the circumference of a circle the law of departure from the circumference and from the centre of the circle is precisely the same as that of a falling body, or of any body acted upon by a given uniform force, the velocities acquired in both cases being in proportion to the time. The tendency, therefore, of a body at *a* to depart from the centre is exactly similar to the action of any force in a given direction in causing a body to move from its initial position. And if a body at *a*, with a motion in the direction of *ad*, is constrained to move in the circumference of the circle toward *e*, as in a groove, or in consequence of being attached to the centre *o* by a cord, corresponding to the radius of the circle, the pressure against the side of the groove in the one case, and the tension of the cord in the other, are the same as the pressure which would arise from the action of any

force on a body of the same mass if it were such as to cause an acceleration in the body, if free to move, equal to that of the rate of departure of the moving body from the centre of the circle. This tendency of a body, in gyrating around a centre, to recede from the centre, and, when not free to depart, to cause pressure in the direction of the radius from the centre, is called *centrifugal force.* It must be understood, however, that this so-called force is not a real force, such as arises from any attracting or propelling force in a given direction, but simply that the gyrating body, if free, would at any given instant be carried away from the centre by its inertia in the varying direction of the radius, in the same manner as it would be moved in a given direction by the action of a constant and real force in this direction. The centrifugal force depends simply upon the inertia of the body, and being always at right angles to the direction of the gyratory motion, it does not increase velocity and momentum, as a real force does, but tends simply to drive a body away from the centre of curvature, and to cause pressure in that direction.

36. The effect of centrifugal force with reference to the varying distance of the body along a radius of constantly varying direction being similar to that of a real force with reference to the motion of the body along a fixed direction, we obtain an expression of this force in the same manner, namely, by multiplying the mass of the body into its rate of departure from the centre at the end of a unit of time, corresponding to what is called acceleration in the case of a real force. In the case of the centrifugal force, however, it must be understood that the unit of time, which is entirely arbitrary, must be so small that the body describes only a very small, strictly infinitely small, arc in comparison with the whole circumference of a circle during this time.

In a falling body, if the unit of time is one second, the velocity at the beginning of the first unit is 0, and at the end of it 32 feet; and as the velocity increases in proportion to the time, the average velocity is 16 feet. Consequently the space fallen through during the first second is 16 feet, one half the

velocity acquired at the end of the second. If now, in Fig. 1, we suppose *ab* to be the space passed over by the body in a unit of time, this unit being such that *ab* is very small in comparison with the radius *ao*, or the whole circumference of the circle, though it is necessarily very much exaggerated as represented in the figure, then *be* is the space over which the body passed during this unit of time in the varying direction of the radius from the centre, or the space by which the distance of the body from the centre is increased. Consequently, from what has just been stated in the case of a falling body, $2be$ is the acceleration of the centrifugal force. Putting, therefore, as usual, m for the mass of the body, we have $2be \times m$ for the expression of the centrifugal force.

But it is desirable to have an expression of this force in terms of the gyratory velocity of the body around the centre of gyration, and the distance of the body from the centre. The gyratory velocity of a body is its velocity, or that component of its velocity, which is at right angles to the radius of gyration. If the body at *a*, Fig. 1, has a velocity which would carry it to *b* in a unit of time, then the distance *ab*, sensibly equal to the arc *ae*, is its gyratory velocity at the point *a*. The gyratory velocity $a'b'$ of any point a' in the radius *ao*, at the distance of unity from the centre *o*, expressed in terms of this unit, is called *the gyratory velocity in terms of the radius*, since this would be the real gyratory velocity if the radius were made the measuring unit. It is the ratio between the gyratory velocity and the radius, since it is the gyratory velocity divided by the radius, and so expressed it is often called simply *angular velocity*, since it is proportional to the rate of change of the angle at the centre. The real gyratory velocity, therefore, is equal to the gyratory velocity in terms of the radius multiplied into the radius. Denoting the gyratory velocity *ab*, Fig. 1, by w, the radius *ao* by r, and the gyratory velocity in terms of the radius by n, we evidently have $ab = ao \times n$, or $w = rn$.

We have in the figure, as has been shown, the angle *bce* equal the angle *aob*; and hence *be* is evidently as many times greater than *ce* as *ab* is greater than *ao*, and as *ab* or w is equal

to rn, be is equal to $ce \times n$, and so $2be$ is equal to $2ce \times n$. But ce is equal to ac or one half of ab, where all are very small, as here supposed, and consequently $2ce$ is equal to ab or w, and hence we have $2ce = w = rn$. Putting rn for $2ce$ in the preceding expression of $2be$, we have $2be = rn \times n = rn^2$, and the preceding expression of the centrifugal force, $2be \times m$, becomes rn^2m. Or putting F_c for the centrifugal force, we have*

$$F_c = rn^2m = \frac{w^2}{r}m.$$

The last form of this expression is obtained from the first one by putting $rn^2 = rn \times n$, and then for rn and n their equals w and w/r. From the last form of expression it is seen that the centrifugal force is directly as the square of the gyratory velocity and inversely as the distance of the body from the centre of gyration.

In obtaining the preceding results it was assumed that the unit of time is very small (strictly it should be infinitely small), but the expressions thus obtained are applicable to any other unit. For velocity expresses simply the relation between an infinitely small portion of space passed over by the body at any instant of time and the corresponding infinitely small portion of time, and this can be expressed by using either large or small units of time, it being understood that the space passed over during the unit of time is that which would result from a continuance of the rate of motion or relation between the infinitely small portions of space and time at any instant through the whole unit of time, however large it may be.

* Joining e and g in the figure, and considering the arc ae a straight line, and be a continuation of the line ge, as we may when they are very small, we have by the well-known proportions of similar triangles $eg : ae :: ae : be$; or putting $ab = ae$, as we can where these quantities are very small, we have $be = \frac{(ab)^2}{eg}$ or $2be = \frac{w^2}{r}$, since where ae is very small we have sensibly $eg = ag = 2r$. Hence, multiplying this expression of the acceleration $2be$ into the mass m, we get the expression of the centrifugal force above.

37. If the body is forced to move in an irregular curve, as in a groove, which is not the circumference of a circle, there is always some circle the circumference of which coincides sensibly with a very small portion of this irregular curve at any point where the body may be moving, and the radius of this circle is called *the radius of curvature.* Since the preceding expression of centrifugal force is deduced from the consideration of only a very small portion of the circumference of the circle, strictly only an infinitely small part, and is dependent simply upon the curvature at that point, it is evident that it is applicable to any point of an irregular curve, if we use for r in the expression the radius of curvature of that point of the curve. Hence, the velocity remaining the same, the centrifugal force is inversely as the radius of curvature.

As an expression of the ratio between the centrifugal force and that of gravity, where both forces act upon the same or equal masses, we get from the preceding expression, by dividing by gm, the force of gravity,

$$\frac{F_c}{gm} = \frac{w^2}{9.806r},$$

in which the units are the second of time and the meter. If the English foot is used as the unit of length instead of the meter, then 32.2 instead of 9.806 must be used as the numerical coefficient of r in the denominator of the expression.

As an example of the practical application of this relation, let us suppose that the body has a velocity on a horizontal plane of 20 meters per second (45 miles per hour nearly*), and that it is constrained to move in a curve with a radius of curvature at any given point of 1000 meters. We then have for the ratio of the centrifugal force to that of gravity at that point $20^2 : (1000 \times 9.806) = 0.0408$. If, therefore, the moving body were that of a railroad car, with a velocity of 20 meters per second, on a part of the road having a radius of curvature of

* To reduce meters per second to miles per hour, multiply by 2.237, and *vice versa.*

1000 meters, the lateral pressure on the side of the rails, in a direction from the centre of curvature, would be about the $\frac{1}{16}$ part of the vertical pressure of the car.

38. If the body is entirely free, and is not constrained to move in a given path, but is acted upon by some centripetal force which is exactly equal to the centrifugal force, it in this case moves concentrically around the centre of the forces in the circumference of a circle. Such a force may be found in the one component of the force of gravity where the body gyrates horizontally upon a surface having a descending gradient on all sides from the exterior part toward the centre of gyration. The *measure* of such a gradient is the ratio between the change of level in a given horizontal distance, and this distance. For instance, if the change of level is one meter in the distance of 1000 meters, the gradient is 0.001, or one in one thousand, what ever the unit of measure. If the body in moving rises to a higher level, it is called an *ascending* gradient, and *vice versa.*

The ratio expressing the gradient is also that between the component of gravity, acting in any direction along the inclined surface, and the whole force of gravity. If de in Fig. 2 is the change of level in the horizontal distance ad, if we denote the ratio expressing the gradient by e we shall have $e = de : ad$. If now we let the vertical line ce represent the force of gravity, and cb is drawn perpendicular to the inclined surface, then, by the principle of the resolution of forces, the whole force of gravity represented by ce is equivalent to two other forces, called components, one of which is represented by the line cb and acts in the direction from c to b, and the other by the line be and acts in the direction from b to e; these components and the whole force being to one another as the sides of the triangle representing them. Denoting the component of the force of gravity represented by be by f, that of the whole force, as we have seen in §9, being represented by gm, we have sensibly, putting $bc = ec$, $f : gm = be : bc$. But the angles at a and c being equal, as is pretty evident without rigorous demonstration, the lines forming the angles must separate from each other in proportion to the distances from the angles, and hence we have

$de : ad = be : bc$. These ratios are also derived more directly from the well-known geometrical relations of similar triangles. Putting the ratios above, which are both equal to the ratio $be : bc$, equal to each other, we get $f : gm = de : ad$; or representing, as above, this latter ratio be e, we get $f : gm = e$, or $f = egm$. Strictly, this expression of f here is the horizontal component of the force down the slope, but in putting $bc = ec$

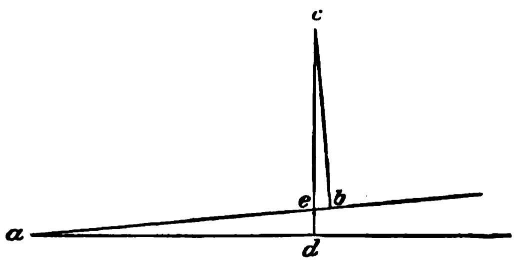

Fig. 2.

above we neglect the very small difference between the two where the gradients are so small.

Where the gradient is very small, as it usually is, the centrifugal force in the direction of the slope is sensibly the same as in a horizontal direction; and so, in order that the centrifugal and centripetal forces on the slope may be exactly equal, the gyratory velocity must be such as to satisfy the condition

$$eg = \frac{w^2}{r} = rn^2,$$

as is readily seen by comparing the expression of the centripetal force above, with that of the centrifugal force F_c in § 36. Since g may be regarded as constant, it is seen from this expression that the gradient of the slope which satisfies this condition varies directly as the square of the gyratory velocity and inversely as the distance r of the body from the centre of gyration. With the same value of w it is seen that the gradient very near the centre, where r is small, would be very steep, and consequently this expression would not hold there, since it has

been assumed above that the gradient is so small that the horizontal and inclined distances do not differ sensibly.

From what precedes, if the railroad car, in the example which we have just had, were to move around upon a surface declining at all points toward the centre of curvature, or, what would be the same, if the outer rail were a little above the level of the inner one, then the centrifugal and centripetal tendencies would exactly counteract each other, and there would be no lateral pressure against the rails, if the gradient between the two rails were determined by the preceding expression. In the example above, the outer rail would have to be higher than the inner one by the $\frac{1}{25}$ part of the width between them. Since the required gradient differs with different velocities, the best that can be done is to adapt the gradient to an average velocity, or rather, to an average of the squares of the usual varying velocities.

Upon the same principle, the gradient between the two banks of a river, flowing in a channel around a centre of curvature, can be determined. If the velocity were 2 meters per second (4.5 miles per hour), and the radius of curvature 1000 meters, then the value of the gradient e from the preceding expression would be $2^2 : (1000 \times 9.806) = .000408$; and hence the level of the water on the opposite side from the centre of curvature would be higher than on the other side the $\frac{1}{2500}$ part, nearly, of the width of the river, if all parts had the same velocity and the same radius of curvature.

39. If a body, as at a, Fig. 1, is at rest relative to the surface, with a descending gradient from all sides toward the centre o, and the whole sub-stratum of the surface gyrates around this centre with a velocity w at the distance r from the centre, it is evident that the centrifugal force F_c and the counteracting component of gravity f along the descending gradient, are precisely the same as if the body gyrated with the same angular velocity around the centre of gyration, and moves without friction over the surface, since the absolute motion of the body is the same in both cases. In order, therefore, that the body may remain at rest upon such a revolving surface and not

move either from or toward the centre, the condition of the expression in § 38 must be satisfied. It is seen from this expression, since n is constant with the same angular velocity of gyration, that the gradient e must be proportional to r, and consequently vanish at the centre where r vanishes and increase as the distance r increases. Such a gradient would be found in a concave surface corresponding to a small segment of a spherical shell, a section of which, *aob*, in which o is the centre, is represented in Fig. 3.

If the whole gyrating stratum were covered by a fluid, confined within a given area by a circular rim, as in the case of a very shallow basin of water revolving horizontally around its centre, then all parts of the surface of the fluid in static equilibrium and at rest relatively to the basin, must have a gradient

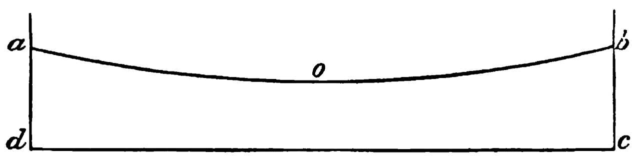

Fig. 3.

satisfying the preceding condition, and be such as that represented in Fig. 3, in which a body at rest upon any part of it would not move either toward or from the centre.

40. If we now suppose the concave surface of the revolving solid stratum, *abcd*, Fig. 3, to be the same as that which a liquid at rest relative to the solid would assume, and that a body placed upon this surface has a gyratory velocity v, either in the same or the contrary direction, the surface gyrating with the velocity ω at the distance r of this body from the centre of gyration, then the absolute gyratory velocity of the body at the distance of r is $\omega + v$, the latter being negative when the body gyrates in the contrary way to that of the solid. Hence, in order to obtain now an expression of the centrifugal force F_c, we must put in the expression of F_c in § 36, $\omega + v$ instead of w. But since the square of the sum or difference of two quantities, or of numbers representing them, is equal to the square of the

first, plus or minus twice the product of the first into the second, plus the square of the second, as any one not familiar with algebraical operations can easily verify in special numerical cases, we shall now have for the expression of the centrifugal force

$$F_c = \frac{(\omega + v)^2}{r}m = \frac{\omega^2 + 2\omega v + v^2}{r}m.$$

The first term of this expression $\omega^2 m/r$ is the part of the centrifugal force which exactly counteracts the tendency of the body to slide down the slope toward the centre of concavity and of gyration, since we have assumed that the gradient of the surface is such as to satisfy the condition of § 38, and consequently such as a liquid surface would assume where all parts have the same angular velocity of gyration. The second part of this expression, containing both ω and v as factors, depends upon both ω and v, and vanishes if either ω or v is equal to 0. If v is the velocity of gyration in the same direction as that of ω, then the effect of this term is positive, and the part of the force represented by it tends to drive the body away from the centre of the slope, being so much force in addition to that which is just sufficient to support the body upon the slope and to keep it from sliding down toward the centre. But if v is negative, that is, contrary to the direction of ω, then the centrifugal force is diminished, and is not sufficient to counteract the tendency of the body to slide down the inclined surface toward the centre. The effect of this term, therefore, is to deflect the body on the concave surface, away from the centre when v is positive, but toward the centre when negative, and in both cases to the right of the direction of the relative motion. The last term in the expression of F_c, depending upon v^2, is simply the centrifugal force belonging to the relative gyratory velocity, in case the solid surface had no gyratory velocity ω. Consequently the effect of this part of the force is always to drive the body from the centre, whether v is positive or negative.

On a gyrating surface, therefore, in which the condition of § 38 is satisfied, namely, that the body when at rest has no tendency to move either toward or from the centre, the gradient and the gyratory velocity being such that these two tendencies exactly counteract each other, we then have for the expression of the effective centrifugal force

$$F_c = \frac{2\omega v + v^2}{r} m.$$

THE PRINCIPLE OF THE PRESERVATION OF AREAS.

41. If a body has a gyratory motion around any given centre, and is also acted upon by a force in a direction either toward or from that centre, the variable line connecting the body with the centre, called the *radius vector*, always sweeps over equal areas in equal times. Thus, if a body at a in the annexed figure has a gyratory component of velocity from left to right around the centre at o, and is at the same time acted upon by a centripetal force of any kind, so as to cause it to move, in successive equal intervals of time, to the points b, c, and d, then the areas aob, boc, cod, etc., are equal. This was demonstrated by Newton, and is a firmly established principle in mechanics. This equality of areas, of course, holds for each area, in any part of the path of the body, swept over by the radius vector in an infinitely small portion of time, and these areas are equal to one half of the product of the radius vector into the component of motion in this infinitely small portion of time which is perpendicular to the radius vector, called the gyratory

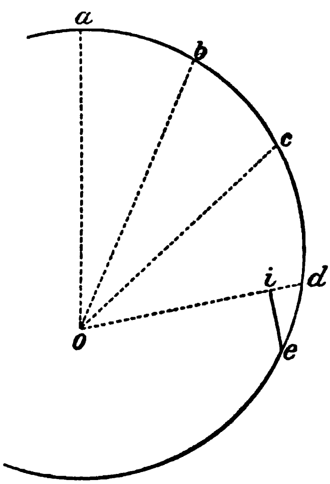

Fig. 4.

component of motion. It also holds where the body recedes from the centre in consequence of either the centrifugal force arising from the gyratory component of motion, or any real force acting in a direction from the centre.

The velocity of a body of variable motion at any point is the space it would pass over in a unit of time, as one second, if the rate of motion which it has at that point should continue through that time, and not the space which it actually passes over in a unit of time with varying velocity. The gyratory component of velocity is that part of its actual velocity in the path described, resolved in a direction perpendicular to the radius vector. For instance, if the body at d in the figure above has a rate of motion which, if unchanged, would carry it to e in the direction of the tangent in a unit of time, then if this velocity de is resolved into the two components ei and di, the former is called the *gyratory* and the latter the *centripetal* velocity. As the velocities depend upon the rate of motions at any given point, and not upon the actual motion in a unit of time, and the areas swept over in infinitely small portions of time are equal to half the product of the radii vectores into the gyratory components of motion, and these areas are all equal, it is evident that the gyratory velocities of the body in different parts of its orbit are inversely as the radii vectores. Putting, therefore, as heretofore, w for the gyratory velocity and r for the radius vector, or variable radius, we have for all parts of the path of motion

$$rw = c,$$

in which c is a constant. This is the expression of the law of equal areas in this case. Hence the nearer the body comes to the centre the greater is the gyratory velocity, and *vice versa.*

A familiar example of this law is observed in twirling a small body attached to the end of a string around the end of a stick, and, after the gyratory motion has been established, allowing the string to wind up around it. The string acts as a force, continually drawing the body closer to the axis or centre of

gyration, and as this is done the gyratory velocity increases, and when the string becomes very short the velocity becomes very great. Just the reverse takes place if the motion is such as to unwind the string, and to allow the body to recede to greater distances from the centre of gyration. As the distance increases, the gyratory velocity becomes less; and when the distance becomes very great, the gyratory velocity becomes very small.

A more perfect exemplification of this law is found in the motions of the planets and comets in their orbits around the sun. In this case the law of the centripetal attractive force is such as to cause *elliptical* orbits; but the law holds, whatever may be the law of this force and the path of the body. At aphelion, where the planet or comet is at its greatest distance, the gyratory velocity is least; but as it comes nearer to the sun in moving around to its perihelion, the gyratory velocity increases in accordance with the law above, except so far as it may be slightly affected by the perturbations of the other planets. In the case of comets with very eccentric elliptical orbits, the radius vector at perihelion becomes very small, and the corresponding gyratory velocity enormous.

42. As the body is drawn or forced toward the centre of gyration, the gyratory velocity, if the body is free, is increased, just as in the case of a force acting upon a body in the direction of its motion, and the amount of acceleration in a unit of time, multiplied into its mass, in the former case as in the latter, is the measure of the force causing the acceleration. If the body is constrained to move in a given path, not directly toward the centre, or is retarded by frictional or other resistances, a part of this force may be spent in causing pressure and overcoming the resistance of the friction, and the balance only, in causing acceleration in the direction of motion.

According to the preceding expression of the law of equal areas, $rw = c$, if r is decreased by the action of any kind of centripetal force by the $1/m$ part of r in a very, strictly infinitely, small portion of time, m being a very large or infinitely great number, then w is increased by the same part, that is, by

$1/m$ part of w, or w/m. If we now suppose that r continues to decrease by the same rate during a unit of time, and we call this rate, or centripetal velocity of the body, x, then r is changing at the rate of the x/r part of it in a unit of time; and consequently w, if it increased uniformly, would increase the x/r part of w or xw/r in a unit of time. Now the increase of velocity in a unit of time is called the acceleration, and the acceleration multiplied into the mass m is the measure of the force which causes the acceleration. And this force acting in the direction of the gyratory velocity w, and so at right angles to the radius, may be called *the gyratory force.* Putting, therefore, F_g for this force, we have

$$F_g = \frac{xw}{r} m.$$

The gyratory force, therefore, is directly as the product of the gyratory velocity w into the centripetal velocity x, and inversely as the distance r of the body from the centre. If r, therefore, could become infinitely small, the gyratory force would become infinitely great. By comparing this expression with that of the centrifugal force F_c, § 36, it is seen that they are both equal where x and w are equal, that is, where the velocity of the body toward the centre is equal to the gyratory velocity.

43. If we suppose that a horizontal plane has a gyratory motion around the centre o in Fig. 5 (p. 58) with a velocity ab at the distance of a from the centre, the line AB being supposed to be fixed in space, and also that a body at a, while acted upon by a centripetal force toward o, also has a gyratory velocity relative to the plane equal to bc, then the absolute gyratory velocity is ac. Denoting now, as in § 40, by ω the gyratory velocity ab of the plane at the distance of oa or r, and by v the gyratory velocity bc relative to the plane, then the absolute gyratory velocity ac of the body is $\omega + v$, and hence in this case, in order to get the gyratory force F_g, we

must put $\omega + v$ for w in the preceding expression, and we thus get

$$F_g = \frac{x\omega}{r}m + \frac{xv}{r}m = \omega + v)\frac{x}{r}m.$$

But the gyratory velocity, reckoned from the line ob instead of the fixed line AB in space, is greater by the difference

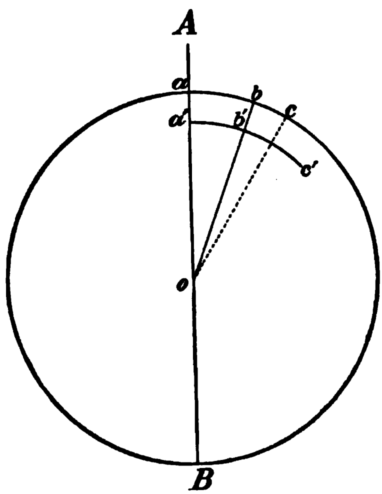

Fig. 5.

between ab and $a'b'$, supposing the body to move with uniform velocity toward the centre by the distance aa' in a unit of time. Now in a unit of time under these conditions ab is greater than $a'b'$ by the x/r part of ω, or $x\omega/r$. Hence if the gyratory velocity is reckoned from the line ob instead of the fixed line AB, we must add $x\omega/r$ to the acceleration where the gyratory velocity is reckoned from the gyrating line ob fixed relatively to the plane, but not in space. With this additional term added to the acceleration in the previous case, we get for the relative gyratory force

$$F_g = \frac{2x\omega}{r}m + \frac{xv}{r}m = (2\omega + v)\frac{x}{r}m.$$

By comparing this expression with the last expression of the centrifugal force F_c in § 40, it is seen that they are the same if x and v are equal, that is, if the centripetal velocity is equal to the relative gyratory velocity. From this it is to be understood that, if a body has a gyratory velocity relative to a gyrating surface around the same centre, in which the condition of § 40 is satisfied, and the gyrating body is drawn in toward the centre by a centripetal force of any kind, with a given velocity, there results a gyratory force which tends to overcome resistances to gyratory motion on the surface, or the inertia of the body, when there are no frictional or other resistances, and to cause accelerated gyratory velocity, and that this force is exactly equal to that which tends to drive the body away from the centre when the gyratory velocity is equal to the centripetal velocity in the other case. If the force is centrifugal instead of centripetal, then the direction of the gyratory force is reversed, and the relative greater velocity is diminished and may be even overcome and reversed as the distance of the body from the centre is increased; just as has been shown, § 40, in the other case when the gyratory velocity v is negative, or in the contrary direction, the deflecting force tends to draw the body from the direction of motion toward instead of from the centre.

If the body is not free to gyrate relatively to the gyrating surface, but is constrained to move either directly towards or from the centre, as in a groove, then the gyratory force causes pressure against the one side or the other, according as the body is forced toward or from the centre.

CENTRIFUGAL FORCE IN MOTIONS ON THE EARTH'S SURFACE.

44. With the preceding preliminary results with regard to centrifugal force in general, §§ 35–40, we are now prepared to investigate the effect of the centrifugal force arising from the earth's rotation on its axis, upon the motions of bodies upon its surface. Let PCE, Fig. 6 (p. 60), represent one quadrant of a

meridional section of the earth turning on its axis, of which P is the pole, C the centre, and E the point in the equator intersected by this plane, and let a be a body at rest on the earth's surface, subject to the action of the centrifugal force arising from the earth's rotation and the consequent gyration of the body around the point e as a centre in the semi-axis PC. The amount of this force is given by the expression of F, in § 36, and it is in the direction of the line ab in the plane of gyration, and the body would be driven away in this direction if this centrifugal force were not counteracted by the stronger force

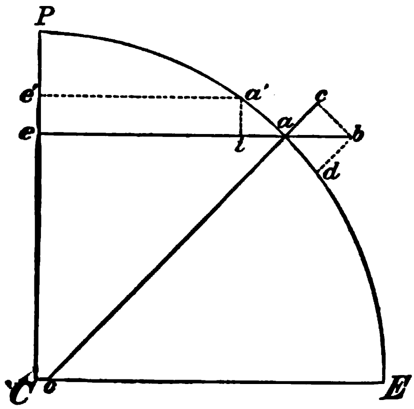

Fig. 6.

of the earth's attraction. Let us now suppose that the centrifugal force is represented by the line ab, and that this is resolved into two component forces in the directions of ad and ac, the former parallel and the latter perpendicular to the earth's surface, which together are equivalent to the force represented by ab. If the points c and d are determined by perpendiculars drawn to these lines from the point b, thus completing the parallelogram of forces, it is well known that the original or resultant force, and the two components in the directions of ad and ac, are respectively proportional to the lines ab, ad, and ac.

The component of the centrifugal force acting in the direction of ad, and represented by this line, tends to drive the body

on the earth's surface from the pole toward the equator, and if the surface were perfectly smooth and without friction, would do so on a spherical globe; but this tendency is exactly counteracted by the tendency to slide down the ellipsoidal surface of the earth toward the pole *P*. The ellipticity of the rotating earth is necessarily such that these two tendencies or forces are exactly balanced, and just as in the case of the concave gyrating surface in § 39, the concavity, and gradient of the slope at each point, are such that the tendency to slide down toward the centre is exactly counteracted by the centrifugal force. In fact the earth's surface of either hemisphere, referred to a spherical surface, or even one everywhere normal to the direction of the earth's attraction, is a concave surface, lower at the pole than at the equator. A body, therefore, at rest upon the ellipsoidal surface has no tendency to move, either toward the equator on account of one component of the centrifugal force, or toward the pole on account of the earth's ellipticity.

The other component of the centrifugal force, acting in the direction of, and represented by, *ac*, tends simply to counteract a comparatively small part of the force of the earth's attraction, thus causing the force of gravity to be a little less than that of the earth's attraction.

45. The centrifugal force of the earth's rotation acting in the plane of gyration is, by the first form of the expression of F_c in § 36,

$$F_c = rn^2m,$$

in which r corresponds to ae in the figure, and in which n is the gyratory velocity of the earth's rotation in terms of the radius. This being a constant, this force is proportional to r, and this latter is proportional to the cosine of the latitude if we neglect, as we may in researches of this kind, quantities of the order of the earth's ellipticity.* Putting, therefore, l for the latitude

* Neglecting such quantities, the line ao, Fig. 6, can always be taken as the mean radius r', and may be supposed to be drawn to the centre C instead of o, and so perpendicular to the surface at a. Then the angle $aCE = CaE$ is the latitude of the body at a, and we consequently have $r' : r = 1 : \cos l$, or $r = r' \cos l$. And as r' is constant, we have $r = \cos l$ when r' is taken as unity.

and r' for the mean radius of the earth, we have $r = r' \cos l$, and hence the preceding expression becomes

$$F_c = r'n^2 \cos l \, m.$$

The cosines for each fifth degree of latitude are given in Table V, to which those who are not familiar with such functions can refer, and from which it is seen that they are greatest at the equator and vanish at the poles. The horizontal component of the centrifugal force, represented by *ad* in the figure, varies on the different latitudes in comparison with the whole, as the sines of the latitudes. These also are given in Table V, from which it is seen that this component, in comparison with the whole, is greatest at the pole and vanishes at the equator. But as the whole varies as the cosines of the latitudes, the horizontal component varies as the product of the cosine into the sine. Putting, therefore, F_h for the horizontal component of the centrifugal force, we have

$$F_h = r'n^2 \cos l \sin l \, m.$$

This component, therefore, vanishes both at the equator and the pole, and is a maximum on the parallel of 45°*.

Since the earth performs a complete sidereal revolution on its axis in 23h. 56m. 4s., or 86,164 seconds, any part of the earth, at the distance of unity from the axis, moves through the space $2\pi = 6.2832$ such units in this time. Hence the space moved through in one second, in terms of the radius, is $6.2832/86164 = 0.00007292$, which is the value of n above. The equatorial radius of the earth being 6,378,190 meters, we get for the acceleration of the centrifugal force at the equator, in meters, or, in other words, the force for unity of mass, $r'n^2 = 6{,}378{,}190 \times (0.00007292)^2 = 0.033913$. Comparing this with

* By a well-known transformation we can put $\cos l \sin l = \frac{1}{2} \sin 2l$. On the parallel of 45° we have $2l = 90°$, and hence the maximum is on that parallel as stated.

the acceleration of the force of standard gravity, which is 9.806 m., we get 0.033913 : 9.806 = 1/289, very nearly.

The velocity of rotation of any point on the earth's surface is $rn = r'n \cos l$, since n is simply the ratio between this velocity and the radius of gyration r. These values, and also the values of r, corrected for the effect of the earth's eccentricity, are given for each degree of latitude in Table V.

Since the centrifugal force is as the square of n, if the gyratory velocity of rotation were seventeen times greater, the centrifugal force at the equator would be very nearly equal to that of the earth's attraction, since the square of 17 is equal to 289, and then a body at the equator would have little or no pressure upon the earth's surface; and if the rotary velocity were still a little more increased, the body would fly off into space,—not, however, in the direction of a tangent, as is often said, but in an elliptic orbit, which would bring it back again to the earth's surface.

46. If a body at a, Fig. 6, has, in addition to the gyratory east velocity due to the earth's rotation, also an easterly* velocity relative to the earth's surface, called relative velocity, the absolute east velocity is then increased by the east component of this velocity, and the centrifugal force in the plane of gyration is increased in proportion to the square of the absolute east velocity. The centrifugal force is then given by the first form of the expression of F_c, in § 40, using for ω and r, for any given latitude, the values in Table V. But the horizontal component of this force, the part which tends to drive the body from the pole toward the equator, we have seen, is to the whole as the sine of the latitude to unity; and hence in order to obtain the expression for the horizontal component F_h in this more general case, in which the body has an east or

* In this work east or eastern is used in an exact sense, so that an east or eastern direction means one exactly toward the east, while easterly is used more vaguely to indicate directions varying considerably from an exact eastern direction, but having a large east component. An east wind is one coming exactly from the east, while an easterly one may have a considerable north or south component of motion. So for all other points of the compass.

west relative velocity v in addition to that of the earth's rotation, we must multiply the expression of F_c, in § 40, into sin l, and we thus get

$$F_h = \frac{\omega^2 + 2\omega v + v^2}{r} \sin l\, m.$$

The first term in this expression $(\omega^2/r) \sin l\, m = rn^2 \sin l\, m = r'n^2 \cos l \sin l\, m$, putting $r = r' \cos l$ to obtain the last form, is the part of the horizontal component of the centrifugal force, as shown in the last section, which is exactly counteracted by the tendency of the body to slide down the elliptical surface toward the pole, so that in the case of a body resting upon such a surface, and having an easterly relative velocity with an east component v, the force which tends to drive it toward the equator is expressed by the last two terms above; so that we get in this case, putting F_v for the force deflecting to the right or south of the direction of motion of the east component of velocity v,

$$F_v = (2\omega + v)\frac{v}{r} \sin l\, m.$$

This is the same as would be obtained by multiplying the last expression of F_c, in § 40, into sin l.

If the relative velocity is westerly, then the west component v is negative; and the absolute velocity of gyration, and consequently the centrifugal force and its horizontal component, are *less* than in the case in which the body is at rest on the earth's surface, and in which case $v = 0$. This force, therefore, in this case, is not sufficient to counteract completely the tendency of the body to move on the ellipsoidal surface toward the pole, and there is a residual part of this tendency or force left, equal to the preceding expression where v is negative, and the body consequently tends to move toward the pole. If, therefore, the body has a relative motion with an east component of motion, it is deflected toward the equator; but if it has a relative motion

with a west component of motion, it is deflected toward the pole; and consequently in both cases to the right of the direction of the east or west component of motion in the northern hemisphere, and the contrary in the southern, where sin l is negative.

If the earth had no rotation on its axis and the body had an east or west component of velocity v, we should still have the last term in the preceding expression of F_h, depending upon v^2, which is independent of ω, the earth's gyratory velocity, and is the same as if the body had an absolute east or west velocity equal to v upon the earth without rotation on its axis. This part of the force, however, is generally very small in all ordinary cases of relative motions on the earth's surface, since the velocities of these are generally very small, in comparison with the velocities of the earth's rotation, as given in Table V, except very near the poles. For instance, on the parallel of 45°, where $\omega = 329$ meters per second, if v had one tenth of this value, say 33 meters (about 74 miles per hour), it is seen from a comparison of the two terms in the preceding expression of F_v, that the latter, which is independent of ω, would be only $\frac{1}{20}$ part of the other depending on ω. If the relative gyratory velocity were west and equal to twice the east velocity of the earth's rotation, then, it is seen, the whole deflecting force F_v would vanish, since then v would be equal to 2ω, and be negative.

THE PRINCIPLE OF EQUAL AREAS IN MOTIONS ON THE EARTH'S SURFACE.

47. If a free body upon the spheroidal surface of the earth, supposed to be perfectly smooth and without friction, has an absolute gyratory velocity around the earth's axis, and is also acted upon by a force in the plane of the meridian in a direction either directly toward or from the pole, then, as the body approaches or recedes from the axis, the principle of equal areas explained in § 41 holds in this case also, just as it does where

the body is forced directly toward or from the centre of gyration, provided we consider the motions of the body, and the areas swept over by the radius, as projected upon a plane perpendicular to the axis of the earth's rotation.

Since the perfectly smooth surface of the earth without friction can have no effect upon the absolute motions of the body in space, they are entirely independent of the earth's rotation, and are the same as if the earth were at rest. If the body on the parallel of a, Fig. 6, has an absolute gyratory velocity along that parallel, either the same as, or greater or less than, that of the earth's surface, and is forced toward the pole until it arrives at the parallel of a', and the radius becomes $a'e'$ instead of ae, the effect upon the absolute gyratory velocity around the earth's axis is the same as if the body had been forced directly in the same plane from a to i, and the absolute gyratory velocities on the two parallels are inversely as the distances from the axis. For the force acting along the earth's surface in the plane of the meridian may be resolved into two components, the one in a direction parallel to the earth's axis, and the other perpendicular to it, so that the former has no effect upon the gyratory velocity, and the latter only is a central force, directed toward the earth's axis and the centre of gyration. The absolute gyratory velocity, therefore, is determined by the equation of § 41, $rw = c$, where the value of c is known. If the values of r and w are known for any given latitude, and are denoted by r' and w' respectively, then we have $r'w' = c$, and the preceding equation becomes

$$rw = r(\omega + v) = r'w',$$

putting, as heretofore, for w the sum of its two components ω and v, the former being the gyratory velocity of the earth's surface and the latter that of the body relative to the earth's surface. From this expression, with the values of r' and w' corresponding to any given latitude l', the value of w, and also of v, are known for any latitude l. For since the values of r

for different latitudes are as the cosines of these latitudes, by dividing each term of the preceding equation by r, we get

$$w = \omega + v = \frac{r'}{r} w' = \frac{\cos l'}{\cos l}(\omega' + v'),$$

in which ω' and v' are the values of ω and v corresponding to $l = l'$. The values of the cosines and also of ω' are given in Table V.

48. The following important applications may be made of the preceding principle by means of this expression. If a body on the equator having no relative east or west velocity were forced toward the pole, the value of $\cos l'$ in this case being unity, we should have, by means of the values of ω and $\cos l$ in Table V, on the parallel of 45°, where $\cos l$ is equal to 0.707, for the value of w, $465/0.707 = 658$ meters per second; on the parallel of 60°, where $\cos l = 0.5$, the value of w would be $465/0.5 = 930$ meters per second; on the parallel of 80°, where $\cos l = 0.174$, we should have $w = 465/0.174 = 2674$ meters per second. Also, for points much nearer the pole it is seen that w would be still very much greater.

But these are the velocities reckoned from a fixed plane in space and not from a given meridian on the earth's surface. To obtain, therefore, the relative east velocities, or values of v, we must subtract from these the values of ω in Table V, which are the absolute east or gyratory velocities of the earth's surface. Hence, deducting these from the absolute velocities above, we get, on the parallel of 45°, $v = 658 - 329 = 329$ meters per second; on the parallel of 60°, $v = 930 - 232 = 698$ meters per second; and on the parallel of 80°, $v = 2674 - 81 = 2593$ meters per second.

If the body before leaving the equator had an east or west relative velocity, then the absolute velocity here, w', would be greater or less by this amount, and so the absolute velocities on the several parallels above, from which the values of ω have to be deducted in order to obtain the relative values v, would all be greater or less in the same proportion.

If the initial position of the body were that of some parallel between the equator and the pole without any initial relative motion, then, if forced *toward* the pole, it would gradually acquire an east component of relative velocity, which, near the pole, would be enormously great, but if forced *from* the pole, it would gradually acquire a west component of velocity relative to the earth's surface. Thus a body without any relative east or west velocity on the parallel of 30°, where the absolute east or gyratory velocity of the earth's surface is 403 meters per second, on arriving at the parallel of 60° would have this velocity increased inversely as the cosines of the latitudes, that is, as 0.500 to 0.866, and hence would be 698 meters per second, and deducting from this the value of ω, in Table V, for this parallel we should have a relative east velocity $v = 698 - 232 = 466$ meters per second. But if the body were forced directly toward the *equator*, on arriving there its absolute east velocity would be decreased inversely as the cosine of 30 degrees to unity or as 1 to 0.866, and hence would be 349 meters per second. Deducting this from the absolute easterly velocity of the earth at the equator, we get for the relative west component of the velocity of the body at the equator $465 - 349 = 116$ meters per second.

If the initial relative west component of velocity were exactly equal to the absolute east or gyratory velocity, either at the equator or any parallel of latitude, then the initial absolute gyratory velocity w' would be 0, and the body then would move in space directly toward or from the pole, and its relative west velocities, at all latitudes, would be those of the absolute east velocities of the earth's surface.

The preceding results, obtained from the principle of § 41, differ very much from those obtained from the principle adopted by Hadley, about the year 1735, in explaining the trade-winds, namely, that a body, being forced directly toward or from the pole, tends to keep its initial absolute gyratory velocity. He says:[7] "A particle of air drawn from the tropics, where it is supposed to have no motion east or west, toward the equator acquires a westward velocity on account of the parallels

continually enlarging. The increase of the parallels from the tropics to the equator is in the ratio of 917 to 1000, and hence the westward motion in an hour is 83 miles at the equator, which is decreased by the effect of the earth's surface to what is observed." But from what has been shown above, the velocity of a body having an absolute east velocity of 917 miles per hour at the tropics, as here supposed, would be decreased at the equator in the ratio of unity to the cosine of the latitude at the tropics, or as 1 to 0.917. Hence, instead of still having an absolute east velocity of 917 miles per hour, it would have a velocity of only $917 \times 0.917 = 841$ miles per hour. This being deducted from 1000, the supposed absolute east velocity of the earth's surface at the equator, we have 159 miles for the relative west velocity there, instead of 83 miles as given above.

According to this erroneous principle, in order to obtain the relative east components of velocity of the body at the several latitudes, in moving from the equator to the pole in the preceding example, it would be necessary to subtract the absolute east velocities of the several parallels, as given in Table V, from that of the equator, and we should thus get in meters per second, at the parallel of 45°, $v = 465 - 329 = 136$; at the parallel of 60°, $v = 465 - 232 = 233$; and at the parallel of 80°, $v = 465 - 81 = 384$.

These differ very much from the velocities obtained above upon the correct principle, which is that of equality of areas in the case of central forces.

49. The gyratory force F_g in this case—that is, the tendency of the body, when free, to acquire an east component of velocity relative to the earth's surface, and when not free, to cause pressure and to overcome resistances to such relative motion—is the same as in the case of any gyrating surface where the body is forced toward the centre with the velocity of x, and is given by the last expression of F_g in § 43. But this may be given in a function of the polar component of velocity of the body in moving on the earth's surface instead of a function of x, which is the velocity, in this case, with which it approaches the earth's axis. From Fig. 6 it is seen

that by similar triangles we have $aa' : ai = aC : Ce = 1 : \sin l$, neglecting quantities of the order of the earth's eccentricity, and supposing aa' to be so small as not to deviate sensibly from a straight line. If we let this represent the polar component of velocity of the body on the earth's surface and denote it by u, then $ai = x$ is the component of this velocity perpendicular to the earth's axis. We therefore have $u : x = 1 : \sin l$, or $x = u \sin l$. Putting this expression of x for x in the last expression of F_g in § 43, we get

$$F_u = (2\omega + v)\frac{u}{r} \sin l\, m,$$

in which, in analogy with the expression of F_v in § 46, F_u is put, instead of F_g, for the force deflecting to the right or east of the direction of motion of the polar component of velocity u.

By comparing this expression with that of F_v in § 46, it is seen that they are very similar; in this a polar component of velocity u gives rise to a force deflecting to the right of the direction of this component, and in the other an east component of velocity v gives rise to a force also deflecting to the right of the direction of this component. And it is also seen that these forces are precisely equal if u and v are equal. When the directions of motion are reversed and u and v become negative, of course the deflecting forces are reversed in their directions of action, but the deflecting force is still toward the right of the directions of these components. This must be understood of the northern hemisphere where $\sin l$ is positive. In the southern hemisphere, where $\sin l$ is negative, of course the deflecting forces are reversed in the directions of their actions, and are always to the left of the directions of the components, whether these are north or south, or east or west.

We have by definition in § 45, $n = \omega/r$; and if, by analogy, we put $\nu = v/r$, we shall have ν equal to the relative gyratory velocity of the body around the earth's axis in terms of the radius, as n is the absolute gyratory velocity, in the same terms, of a body fixed on the earth's surface. Substituting ν

and v for their equivalents in the preceding expressions of F_u and F_v, we get

$$F_u = (2n + v)u \sin l\, m;$$
$$F_v = (2n + v)v \sin l\, m.$$

RESULTANTS OF THE TWO FORCES AND MOTIONS.

50. Where the motion of a body upon the earth's surface is such as to change both its longitude and latitude at the same time, both of the preceding forces are called into play at the same time, and they give rise to one resultant force in a certain determinate direction. If, in Fig. 7, we let AB represent the

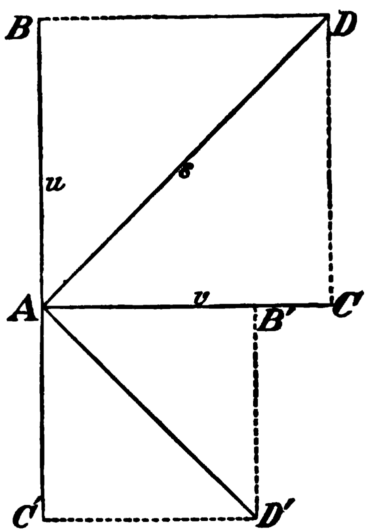

Fig. 7.

velocity u and direction of the polar component of motion, and AC, at right angles to AB, the east component of velocity; then AD, the diagonal of the parallelogram, which we shall denote by s, represents the resultant velocity. But the forces arising from the component velocities, which act in directions toward the right and at right angles to the directions of motion, and consequently at right angles to each other, may be

represented by the lines AB' and AC', drawn perpendicular to, and to the right of, the directions of motion, and from what is shown in the preceding expressions of F_u and F_v, they are proportional to the velocities u and v respectively. Completing the parallelogram, the diagonal AD' then represents the resultant of the two forces, both in magnitude and direction, and as the forces represented by AB' and AC' have been denoted by F_u and F_v respectively, we may let F_s denote, by analogy, the resultant force represented by AD', since it depends upon the resultant velocity s, just as the others do upon the component velocities u and v. Now by a well-known theorem we have the diagonal of a parallelogram equal to the square root of the sum of the squares of two contiguous sides. Hence we have

$$s = \sqrt{u^2 + v^2}, \quad \text{and} \quad F_s = \sqrt{F_u^2 + F_v^2}.$$

If we now square both members of the last two equations of § 49 and take the square roots of their sums, we get, by putting s and F_s for their equivalents above,

$$F_s = (2n + \nu)s \sin l\, m.$$

Since, by the construction of the figure and the relations of the homologous sides of similar triangles, we have $AB : AC (= BD) = AB' : AC' (= B'D')$, and so the angle BAD equal the angle $B'AD'$, by adding the angle CAD to both, we have the angle BAC equal the angle DAD', and hence the latter is a right angle. The direction, therefore, of the resultant of the two deflecting components is at right angles to the direction of the resultant motions, of which the velocity is represented by the line AD or s. As the components of velocity u and v may have any value, either positive or negative, the resultant motion with velocity s may have any direction, and the resultant deflecting force will always be at right angles to, and to the right of, this direction in the northern hemisphere, where $\sin l$ is positive, but to the left in the southern hemisphere, where $\sin l$ is negative, and where, consequently, the resultant force

F_s is reversed in direction. This deflecting force, however, is not a real force, but is of the same nature as centrifugal force, § 35; and as it always acts, as we have just seen, in a direction at right angles to the direction of motion, its tendency is to continually change direction only, and not to increase or decrease velocity and momentum. But this change of direction must not be understood to be a change of absolute direction in space, for this would require a real force, but simply a change of direction relative to the earth's surface.

51. In obtaining the preceding expression of F_s, the only condition with regard to any real force to which the body may be subject was that it shall act only in the plane of the meridian, and so one component of the force be a central force, directed toward the earth's axis. The initial values of u and v, and so of s, are entirely arbitrary, and the value of the latter is only changed subsequently by the action of the real force acting upon the body, in a direction either toward or from the pole. We consequently have no relation between the force and the value of s, and the value of the preceding expression at any time depends simply upon that of s, and in no way upon the force acting upon the body by which the value of s is changed. The preceding expression of F_s, therefore, holds when the force is infinitely small, and so when it entirely vanishes and the value of s and the direction of motion depend upon an instantaneous impulse. This expression, therefore, holds wherever the body has a motion in any direction relative to the earth's surface, in whatever way this motion may have been produced. If a body, therefore, were set in motion upon the spheroidal surface of the earth, supposed to be perfectly smooth and without friction, it would continue to move on forever with its initial velocity, but with continually changing direction relative to the earth's surface, and its path would vary very much with different initial velocities. It is seen from the expression of F_s that the deflecting force, for the same value of s, would differ with different directions of motion, since $\nu = v/r$ is a function of v, which is the east component of s, and so changes with different directions, and even becomes

negative when v becomes a west component, as it may in many cases even where the initial direction is east. It is also seen that this force varies with latitude, so that if the initial velocity given were great, the body might be carried from a high latitude down to a very low one, when sin l and consequently the whole expression would be small, or it might even be carried across the equator into the opposite hemisphere, where the deflecting force would be changed from the right-hand to the left-hand side of the direction, or the contrary, according to the hemisphere from which it passes.

In case the earth had no rotation around its axis, n in the preceding expression of F_s would vanish, and it would then be the expression of the deflecting force due simply to the centrifugal force of the motion of the body. But it must be here borne in mind that these so-called forces, as they have been defined and used, are not real forces, but simply tendencies to cause departures from the parallels and meridians. If the initial motion is in the direction of the meridian, then v, and consequently $\nu = v/r$, vanishes, and so the expression of F_s vanishes, and consequently there is no tendency of the body to depart to either side of the meridian. But if the initial motion is in the direction of a parallel of latitude, then we have $s = v$, and the expression then simply becomes that of the horizontal component of the centrifugal force of the body in its motion along the parallel, and F_s is then the force with which the body tends to depart from this parallel, and of the lateral pressure of the body in a direction from the pole toward the equator if the bódy were constrained to move in a groove on a given parallel. Since in this case s becomes the same as v, the expression has the same sign and the force the same direction of action, whether the initial motion is east or west. At the equator, where sin $l = 0$, this force vanishes, as in the case where the initial motion is in the direction of the meridian, and the body moves in the great circle of the equator.

WHERE THE CENTRE OF FORCE IS NOT THE POLE.

52. In what precedes, it has been supposed that the body is not acted upon by any real force tending to move it out of the plane of the meridian. We come now to consider the case in which there is some force tending to drive the body either toward or from some point on the earth's surface which is not one of the poles, and for the sake of simplicity we shall here limit the case to a portion of the earth's surface around this centre which is not so great that it cannot be regarded, without material error, as a plane surface. Let us, by way of introduction to what is to follow, consider first the case of a limited area around the north pole. In this case if the area does not extend, say, more than ten degrees from the pole, then we can regard $\sin l$ for this whole area as being equal to unity, without any material error arising from such an assumption; and then we get from the expressions of F_u and F_v, § 49, by putting $\sin l = 1$,

$$F_u = (2n + \nu)u \,.\, m;$$
$$F_v = (2n + \nu)v \,.\, m.$$

We have here the case of the gyrating concave disk of § 39, the concavity being that arising from the spheroidicity of the earth's surface, and being such that the tendency of a body at rest to slide down toward the pole is exactly counteracted by the centrifugal force due to the earth's rotation.

The value of n in this case is that of the earth's rotation on its axis. If the centre of the area considered were on the equator, there would be no gyratory motion of this area around the centre, though of course the centre and the whole area would gyrate around the earth's axis, but not around any axis in the plane of the equator. In this case, therefore, we have $n = 0$. It is reasonable to suppose, therefore, that the value of n is the greatest at the pole, and decreases with decrease of latitude, according to some function of the latitude, from the

pole to the equator. This function, it is well known, is the sine of the latitude. Assuming this here without attempting to demonstrate it, if we put n' for the gyratory velocity in terms of the radius around any point on the parallel of latitude l, we shall have

$$n' = n \sin l.$$

Putting n' or $n \sin l$ instead of n in the preceding expressions, they become

$$F_u = (2n \sin l + \nu)u \,.\, m,$$
$$F_v = (2n \sin l + \nu)v \,.\, m,$$

in which u is the velocity toward the centre and v is the gyratory velocity around that centre, and $\nu = v/r$, in which r is the distance from the centre. The first of these is the expression of the gyratory force, and corresponds to the last expression of F_g in § 43; and the second is the expression of force from the centre, and corresponds to the last expression of F_c in § 40, since $n \sin l = n'$ is the angular gyratory velocity around the centre in this case.

The resultant of these, taken as in § 50, is

$$F_s = (2n \sin l + \nu)s \,.\, m.$$

As in the preceding case, this expression varies with different directions of motion of velocity s, since $\nu = v/r$ varies as v varies, and this being one of the components of s changes with change of direction for the same value of s, and may even become negative—as in the case in which the relative gyratory velocity is in a direction contrary to that of the area under consideration, depending upon the earth's rotation.

As is usual in the case of a central force, the value of $\nu = v/r$ may become very great in comparison with $2n$ or $2n \sin l$; for not only does the value of v become very great near the centre, but r becomes very small, and so for both reasons ν becomes large near the centre. In such cases the motion of the body becomes mostly a rapid gyration around the centre, and the deflecting force is almost entirely the cen-

trifugal force arising from the relative gyratory velocity of the body, the part depending upon n in such cases being very small in comparison with that depending upon $\nu = v/r$.

THE DEFLECTING FORCE OF THE EARTH'S ROTATION.

53. We have seen from what precedes that wherever a body upon the earth's surface has a motion relative to that surface, however produced, there is a deflecting force, depending in part upon the earth's rotation, and partly upon the velocity and direction of motion relative to the earth's surface, and entirely independent of any real forces, by which the body is continually acted upon in a direction at right angles to the direction of motion and deflected from its course, and that this, in the case of central forces, may become very great near the centre of force, where the value of r, the distance from the centre, is small. In all straight-lined motions on the earth's surface, the value of r in the expression of $\nu = v/r$ becomes infinite, and so the value of ν in the preceding expression of F_s vanishes, and the whole deflecting force depends upon the earth's rotation. In the case also in which the body is not acted upon by a local force but is subject only to a force acting in the plane of the meridian, the value of ν in the expression of F_s, § 50, is very small in comparison with $2n$, unless it should be very near the pole, where r, in the expression of $\nu = v/r$, is the distance from the earth's axis. It is seen from Table V that the value of 2ω on the parallel of 45° is 658 meters per second; so for a relative easterly velocity v of any body of only 20 meters per second (45 miles per hour), the ratio of ν to $2n \sin l$ in the expression of F_s is as 20 to 658, since $\nu = v/r$ and $2n \sin l = 2\omega/r$. For any ordinary velocity, therefore, on the earth's surface where the body is not subject to the action of a strong local central force, the whole deflecting force depends mostly upon the influence of the earth's rotation, and is almost entirely independent of the centrifugal force arising from the relative motion on the earth's surface, which would exist in case the earth had no rotation on its axis.

For the part of the deflecting force depending only upon the earth's rotation, we obtain either from the expression of F_s in § 50, or the last one in § 52, by neglecting the part independent of n,

$$F_s = 2ns \sin l\, m.$$

In the northern hemisphere this force acts in a direction at right angles to the direction of motion, and to the right; but in the southern hemisphere, where sin l is negative, the direction is reversed, and consequently at right angles to the left. It is seen from the preceding expression that this force, for any given mass, depends simply upon the sine of the latitude and the velocity s of the relative motion, and is entirely independent of the direction of this motion. Hence, *if a body moves in any direction upon the earth's surface, there is a deflecting force arising from the earth's rotation, which deflects it to the right in the northern hemisphere, but to the left in the southern hemisphere.**

54. This important part of the subject may be viewed in a different manner, and one by which it may perhaps be more readily understood by many readers, than by the preceding method of considering the matter. If a body at a, Fig. 8, has

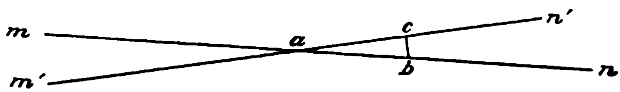

Fig. 8.

a motion with uniform velocity in the direction from m to n, or the contrary, in the line mn, which is continually and uniformly changing its direction from right to left of the direction of motion, it is readily seen that as the line changes its direction, the body by virtue of its inertia has a tendency to continue on in the same direction, and to depart from the gyrating line to the

* This important proposition was first demonstrated by the present writer in the year 1859 in a paper on "The Motions of Fluids and Solids on the Earth's Surface," published in Runkle's *Mathematical Monthly*. This part of that paper appeared in the May number of the Monthly. The whole paper has since been republished by the Signal Service (Professional Paper No. VIII), with extensive notes by Professor Frank Waldo.

right, just as it would from a line fixed in space if there was a constant real force acting on the body, at right angles to the direction of motion. If the body moved in the contrary way from n towards m, the tendency would be to depart on the other side, but still to the right of the direction of motion, where the line gyrates from right to left, looking in the direction of motion. It is also evident that if the line is changing its direction from left to right, the body tends to depart from this line to the left-hand side. But if the body were not free to move in a fixed direction in space, but were compelled to move in the line continually and uniformly changing its direction, as in a groove, then, instead of departing to the right- or the left-hand side, as the case may be, it would press against the side with a force sensibly equal to a real force which, during a very small interval of time, would cause the body to depart from a fixed line in space at the same rate. It is readily seen that this rate of departure at any instant of time must be proportional to the velocity s in the direction of motion, and the rate of change of direction, or the gyratory velocity of the line in terms of the radius. Denoting this latter by n', it is, therefore, proportional to $n's$, and consequently the lateral pressure, where the body is constrained to move in a straight line with changing direction in space, is in the same proportion.

Some general idea of this deflecting force may be formed from the experience of any one in walking over a narrow drawbridge while it is turning around its central pivot. If there were no railings, the tendency would be, if not guarded against, to run off on the one side or the other, according to the direction of gyration; and where there are railings, to press against that side. And this tendency would evidently be in proportion to the velocity of transit across the bridge and the angular velocity of its gyration. Of precisely the same nature is the deflecting force of the earth's rotation; for every horizontal line, fixed relatively to the earth's surface, except at the equator, is continually changing its direction with reference to a direction fixed in space; and so a body, set in motion in any direction relative to the earth's surface, tends, if free, to depart from

this direction, and if constrained to move as in a groove in this direction, to press toward one side or the other, as the case may be.

55. The absolute amount of deflecting force may be deduced in the following manner: If a body at a, Fig. 8, has, at any instant of time, a motion in the direction from m toward n in the line mn with the velocity s, and this line at the same time is changing its direction or gyrating around the point a, whatever other motion the whole line may have, with a gyratory velocity n' in terms of the radius, then if ab represents s, and the angle nan' the change of direction in a unit of time, this unit, which is entirely arbitrary, being taken so small that this angle is very small and ab is sensibly equal to ac, bc being perpendicular to $m'n'$ and measuring the nearest distance of the point b from $m'n'$, the body at the end of this unit of time is departing from the line $m'n'$ by the space bc in a unit of time by virtue of its velocity s in the direction of mn. But this perpendicular at the distance of unity from a is n', and therefore the body in moving through the space s in a unit of time departs from the line $m'n'$, supposed to remain fixed for the time, by the space $n's$. If we now suppose that when the body arrives at b it remains stationary for the time, then the line $m'n'$, by virtue of its change of direction, or gyratory velocity n' around the point a at the distance of unity, is departing from the body at b, which is at the distance s from a, at the rate of ns. Therefore by virtue of both the motions of the body in its direction mn, fixed in space, and also of the uniform change of direction of the line $m'n'$, the body at the end of the unit of time is departing from the line of varying direction in space at the rate of $2n's$ in a unit of time. It is therefore departing from it at the same rate that a body moving in a direction fixed in space would be drawn out of the line of this direction by a force acting in a direction at right angles to the direction of motion, if the acceleration of this force were $2n's$. The lateral pressure, therefore, of a body of the mass m, if constrained to move in the line of varying direction, is $2n's \,.\, m$. In case of a straight line on the earth's surface, at any given parallel l, we

have seen that the value of n' is n sin l. Denoting, therefore, as before, this deflecting force of a body, of velocity s, by F_s, and putting n sin l for n', we have, as in § 53,

$$F_s = 2ns \sin l \,.\, m.$$

In the southern hemisphere the line $m'n'$ would gyrate, with reference to the line mn fixed in space, from left to right; and so in this hemisphere the deflecting force, or tendency to depart from the line $m'n'$, is to the left-hand side.

56. If the body in motion were a stratum of liquid, as the water of a wide river, moving with a velocity s in any uniform direction, then the deflecting force at right angles to this direction would cause an ascending gradient of the surface and of all strata of equal pressure in that direction, and this, as explained in the case of the gradient arising from centrifugal force, § 38, would be

$$e = \frac{F_s}{gm} = \frac{2ns \sin l}{g} = 0.00001487 s \sin l,$$

putting for n and g their numerical values in § 45 in obtaining the last form of this expression.

The value of e being the change of level in the distance of unity, which in this expression is the meter, for a distance of r meters the change of level is re. If a river one mile in width (1609 m.) has a velocity of 2 meters per second (4.5 miles per hour) in any uniform direction on the parallel of 45°, then the surface of the right-hand side in the northern hemisphere stands higher than the other by

$$re = 0.00001487 \times 1609 \times 2 \times 0.707 = 0.034m = 1.3 \text{ inches},$$

in which 0.707 is sin 45°, taken from Table V or any table of natural sines.

The values of e/s have been computed from the formula above, for each fifth parallel of latitude, and given in Table V, from which the values of re can be more conveniently obtained. Thus from the table we get for the parallel of 45°, e/s equal to

0.00001052; and so in the example above, we have, as just obtained,

$$re = 2 \times 1609 \times 0.00001052 = 0.034^{\text{m}}.$$

The value of e above, since it is the ratio between the deflecting force and that of gravity, expresses also the ratio between the lateral pressure, due to the earth's rotation, of a body moving in any uniform direction with velocity s, and the vertical pressure of the body. Multiplying, therefore, the value of e/s in the table by s, we get this ratio. Hence if a railroad car on the parallel of 40° moves with a velocity of 20 meters per second (45 miles per hour), we get

$$e = 0.00000956 \times 20 = 0.0001912.$$

The lateral pressure is therefore about 1/5230 part of the vertical pressure of the car. If the road had a curvature of radius r, then the pressure arising from the centrifugal force, as given in § 37, would have to be added, in the northern hemisphere, if the motion of the car was from right to left around the centre of curvature, but subtracted if the contrary way, in order to obtain the pressure to the right of the direction of motion. It is to be understood of course that the two rails of the road are on the same level.

For intermediate latitudes of the five-degree intervals of the table the value of e/s is readily obtained by interpolation with sufficient accuracy by using only the first differences. Thus, if the value of e is required, for the latitude of 52°, corresponding to a wind velocity of 15 meters per second, we readily obtain from the table, by adding 2/5 of 80 to 0.00001139, $e/s =$ 0.00001171. Hence we have

$$e = 0.00001171 \times 15 = 0.0001756.$$

57. The value of e, for any given velocity s, is the difference of level of the surface of any inelastic fluid, or of a stratum of equal pressure in the case of the atmosphere, between two

stations at the distance of unity. Or, e is the height of the column of the fluid, subject to the force of gravity, which exactly counterpoises a horizontal column of a unit in length and of the same base, subject to the action of the deflecting force causing the gradient. The height of this column is a measure of the difference of pressure on a horizontal surface at the distance of unity, and the vertical pressure of such a column of unit base would express the difference of pressure on such a base in kilograms or pounds of pressure according to the measure adopted. In the case of the atmosphere, however, the gradient is usually measured by the height of the mercurial column which exactly counterpoises the lateral pressure on the same base, § 10, and this is called the *barometric gradient.* But in expressing this gradient we do not use the meter as the unit of length, but use the millimeter in the vertical height of the mercury and one degree of a great circle, or 111,111 meters, as the unit of horizontal distance. The height, also, of the mercurial column is less than that of the air column of standard pressure and temperature, in the ratio of the density of such an air column to that of mercury, or as 1 to 10,517. In order, therefore, to obtain the barometric gradient, which we shall denote by G, from the preceding general expression of e, it is necessary to multiply this expression by 1000, on account of a change of the unit of the vertical column from the meter to the millimeter, and also by 111,111 on account of the change of the unit of horizontal distance, and to divide by 10,517 on account of the height of the mercurial column being less than that of the counterpoising air column in the ratio of 1 to 10,517. We then get

$$G_1 = 0.1571 s \sin l.$$

But it must be understood that this is only for air of standard pressure and temperature. For any other it must be increased or diminished directly as the pressure of the air and inversely as the absolute temperature, since the density of the air by the laws of Boyle and of Charles, § 4, changes in these ratios. Putting, therefore, as heretofore, P_0 for the standard pressure P, and T_0

for the standard absolute temperature T, we obtain the following more general expression:

$$G_1 = 0.1571s \sin l \frac{P}{P_1} \cdot \frac{T_1}{T}.$$

But this expression holds only in the case of straight-lined winds with little or no friction, where the direction of gradient is at right angles to that of the wind.

The preceding less general expression of G_1 gives values at the earth's surface and for all ordinary temperatures, which are sufficiently accurate for most purposes, and these are given in Table V for unit of velocity. For other velocities it is only necessary to multiply the tabular numbers into the velocity. Thus on the parallel of 40°, for a wind of 15 meters per second, we get in millimeters

$$G = 0.1010 \times 15 = 1.51.$$

By reversing the preceding expressions, the value of s, the velocity of the wind, may be found corresponding to any given gradient. By the preceding example, if there is an observed gradient of 1.5 mm., the corresponding wind velocity in a direction at right angles to that of the gradient must be about 15 meters per second.

All the preceding gradients depend upon n, and so upon the earth's rotation, and would vanish if the earth were at rest. In addition to these there may be gradients depending upon the centrifugal force when the wind blows in circuits, and these may be very great if the radius of curvature is small and the velocity large.

In order to obtain the gradient G in the case of centrifugal force in connection with that of the earth's rotation, we must use the expression of F_1, § 52, instead of the preceding one of F_1. We thus get, by multiplying by 1000, by 111,111, and dividing by 10,517,

$$G_1 = \frac{2n \sin l + \nu}{9.806} v \times \frac{111\ 111\ 111}{10517} = 1077.4v\,(2n \sin l + \nu).$$

By neglecting ν, the part depending upon the centrifugal force, and giving n its value 0.00007292, we get the first of the two preceding expressions of G, in the case of standard temperature and pressure. The general expression of this, for any pressure P and absolute temperature T, becomes

$$G_4 = 1077.4v\,(2n \sin l + \nu)\,\frac{P}{P_0}\cdot\frac{T_0}{T},$$

in which $\nu = v/r$.

The relation of the part of the gradient, whether the linear gradient e or the barometric gradient G, depending upon the curvature of the path, to that depending upon the earth's rotation is given by the relation between $\nu = v/r$ and $2n \sin l$ in the expressions of the deflecting forces in § 52. Thus, if a body on the parallel of 45° moves with a velocity of 20 meters per second, and has a radius of curvature of 400 kilometers (400000 meters, or 250 miles nearly) we have $v/r = 20/400000 = 0.00005$, while the value of $2n \sin l$, from Table V, is 0.000103; and therefore the tabular gradient must be increased in the ratio of the sum of these quantities to the latter, or nearly one half, for the effect of curvature and its consequent centrifugal force.

58. If a body has a motion of uniform velocity in any direction, however produced, over the earth's surface, supposed to be entirely smooth and without friction, and the velocity is such that the range of motion in latitude is so small that $\sin l$ can be regarded as a constant, then the lateral deflecting force is constant, and the body is being continually deflected from its course by equal angles in equal times, and so it describes a circle. The radius r of this circle must be such that the centrifugal force corresponding to the velocity s, or $(s^2/r)m$, § 36, in a direction from the centre must be exactly equal to the deflecting force in the contrary direction. Hence, comparing this with the expression of F_s, § 55, we have $s^2/r = 2ns \sin l$, or

$$r = \frac{s}{2n \sin l} = 6857 \frac{s}{\sin l}.$$

Since $\sin l = 0$ at the equator, r would be infinite there, and consequently the body would move sensibly in a straight line so long as it remained near the equator. But if, on the parallel of 45°, where $\sin l = 0.707$, the body should receive a velocity s in any direction of 5 meters per second, we should then have

$$r = 6857 \times \frac{5}{0.707} = 48500 \text{ m.} = 30 \text{ miles nearly.}$$

This would give a range of less than one degree in latitude, and consequently $\sin l$ would remain nearly constant.

59. The acceleration of the deflecting force F_s, § 55, is $2ns \sin l$. In the case of a falling body the space passed over is given by the well-known expression $\frac{1}{2}gt^2$, in which g is the acceleration of gravity, and t is the time. So in the case of this deflecting force, putting $2ns \sin l$ for g, and d for the departure of the body from the tangent or initial direction, we get for a very short range

$$d = 0.00007292s \sin l \,.\, t^2.$$

If a rifle-ball on the parallel of 50° is discharged at a target at the distance of one kilometer with a velocity of 500 meters per second, we have in this case $t = 2$ seconds, and hence we get for the departure d of the ball from the direction of initial discharge relative to the earth's surface

$$d = 0.00007292 \times 500 \times 0.766 \times 2^2 = 0.11 \text{ meter.}$$

The ball, therefore, in the northern hemisphere, would deviate to the right of the target about four inches. It would continue on in the same direction in space, but by the time of its arrival at the distance of the target this would have moved around from right to left about four inches. This lateral deviation,

however, would be very small in comparison with the vertical deviation downward, due to gravity. The ratio between the two is that of the corresponding forces, or of $2ns \sin l$ to g. This, in the preceding example, is

$$2 \times 0.00007292 \times 500 \times 0.766/9.806 = 0.0057,$$

and hence the lateral deviation is only about 1/180 of the vertical one. If, however, the velocity s is increased, this ratio is increased in the same proportion.

The preceding general expression of d is applicable where the velocity of the projectile is variable, provided s represents its mean velocity, just as in the case of a falling body the expression $\frac{1}{2}gt^2$ would be applicable if g were variable, provided we should use the mean of g for all the increments of time. Letting, therefore, D represent the distance to which the projectile is cast, we can put $D = st$, and then the expression of d becomes

$$d = 0.00007292 \sin lDt.$$

This is applicable to a projectile thrown high up in the air, and describing approximately a parabolic curve, in which the horizontal velocity varies. It is only necessary to know the time and distance. If a cannon-ball were projected to the distance of 5 kilometers in 10 seconds on the parallel of 40°, we should have for the lateral deviation

$$d = 0.00007292 \times 0.643 \times 5000 \times 10 = 2.35 \text{ meters.}$$

It is seen that the deviations due to the earth's rotation are of little importance in gunnery; but still it is interesting and important to know of about what order these effects are in ordinary cases.

60. In the case of a projectile thrown vertically upward, the condition of $rw = c$, § 41, must be satisfied, so that the higher the projectile and the greater the value of r is, the less must be the gyratory velocity w. The initial value of this at the earth's surface for the different latitudes is that of ω in Table V, and

the initial value of $r = r' \cos l$ is the initial distance of the body from the earth's axis. The approximate value of r' is 6370 kilometers. Hence we shall have $rw = 6370\omega \cos l$, in which r is the distance of the projectile from the earth's axis. If h is the height of the projectile above the earth's surface, we have $r = (6370 + h) \cos l$. Hence with this we get from the preceding equation

$$w = \frac{6370}{6370 + h}\omega,$$

in which h is expressed in kilometers, but w and ω may have any unit of measure.

At the equator, where $\omega = 465$ m., we have at the height of 10 kilometers

$$w = \frac{6370}{6380} \times 465 = 464.27 \text{ m.}$$

Hence the absolute gyratory or east velocity per second at the altitude of 10 kilometers would be 0.73 m. less than that of the earth's surface, and hence there would be a west velocity relative to the earth's surface of 0.73 meters per second, while at lower altitudes the rate of gain of the absolute easterly motion of the earth's surface upon that of the projectile would be proportionally less. The projectile, therefore, after falling back to the earth's surface, strikes it at a point west of the point from which it was projected, since while at higher altitudes its absolute easterly velocity was less than that of the point of the earth's surface from which it started. The distance between the two points, of course, depends upon the height to which the projectile ascends, and the time between leaving and returning to the earth's surface.

For other latitudes, it is seen from the preceding expression, w is less in proportion to the values of ω in Table V, or in the ratio of the cosines of the latitudes.

CHAPTER III.

THE GENERAL CIRCULATION OF THE ATMOSPHERE.

INTRODUCTION.

61. THE motions of the atmosphere depend, either directly or indirectly, upon differences of temperature. Without these the aqueous vapor would be uniformly distributed in all parts, there would be everywhere the same density, and a perfect calm in the atmosphere would exist over all parts of the globe. The great disturber of uniformity of temperature on the earth's surface is the unequal distribution of the sun's radiated heat. This gives rise not only to differences of temperature, but also to differences in the amount of aqueous vapor, in the air, from both of which causes the density of the atmosphere differs in different places, and thus atmospheric currents are produced. The forces, therefore, which overcome the inertia of the atmosphere when at rest and set it in motion, and which overcome the frictional and other resistances and maintain this motion, depend upon solar heat energy.

Before entering upon the subject of the general circulation of the atmosphere—which, we have seen, is exceedingly elastic and, so far as we know, has no definite superior limit or surface—it will be best to consider a few simpler cases of the motions of inelastic fluids.

If any kind of liquid in a canal, or reservoir of any kind, or the ocean covering the greater part of the earth, had the same density in all parts and were at rest, the surface of the liquid in this case would be everywhere perpendicular to the direction of the force of gravity, and consequently coincide with, or be parallel to, the spheroidal surface of the earth, and the pressures in a horizontal direction on each side of any part of the fluid would be exactly equal, so that it would have no ten-

dency to move in any direction. But if, from any cause, the fluid is so disturbed that its surface is not a level surface, then the pressure of the fluid beneath, at different points of the same level, is different, and horizontal motion takes place in the direction of least pressure. From the nature of a fluid the pressure at any point is equal in all directions, and is not in the direction, only, in which the force acts, as in the case of a solid.

62. If the fluid has the same density in all parts, but is so disturbed, from some cause, from its state of static equilibrium, that its surface is not a level surface, then the fluid at all depths has the same tendency to flow toward the lowest surface level. Let *ABCD*, Fig. 1, represent a part of a longitudinal and vertical section of a canal, and *abcd* that of a cubic unit of the fluid—a cubic centimeter or a cubic inch, for instance; then the pressure on each point of the vertical sides

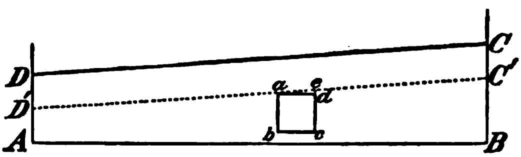

Fig. 1.

of the cube is proportional to the depth of the point below the surface, and the pressure on any one side is equal to the sum of the pressures upon all the differential elements of the surface. If we draw the line *C'D'* parallel to *CD*, it will represent a line in a plane parallel to that of the surface, and hence every point in this plane will be subject to the same pressure. The differences of these pressures, therefore, upon all the differential elements on the same level of the sides of the cube in the directions of greatest and least pressure are equal to the vertical pressure of a column of fluid of the same infinitely small base and altitude *de*, since this is the difference of depth below the line of equal pressure *C'D'* between the corresponding differential elements of the two sides at the same depth. The difference of pressure, therefore, upon the

two sides of the cube in the direction of greatest and least pressure is equal to the vertical pressure of a stratum of fluid of unit base and depth *de*. But the pressure of the whole cube in the direction of the force of gravity is to that of the thin stratum of the same base as their respective altitudes, that is, as *cd* or *ad* to *de*. The force, therefore, which tends to move the cube horizontally in the direction of least pressure, and which is the difference of the pressures on the two sides, is to the force of gravity of this cube as *de* to *ad*; and hence we have the force which tends to move a cube of unit mass horizontally in the direction of least pressure equal to $g(de/ad)$, or to g tan *dae*. But tan *dae* is the change of level in the horizontal distance of unity, or gradient of the inclined surface, which we have denoted by e, § 38. The horizontal force, therefore, for unit mass, becomes, as in § 38, ge, and it is evident that this is the same for all depths, since the preceding reasoning is entirely independent of the depth at which the line $C'D'$ is drawn.

The horizontal force in this case being the same at all depths, the fluid at all depths in the canal has the same tendency to move in the direction of least pressure, and in the case of no resistances from the bottom or sides, all parts in the same vertical line would acquire the same velocity of horizontal motion in a given time, and this would be accelerated until the surface gradient would vanish and become reversed, after which the force arising from the reversed gradient would gradually overcome the momentum and stop motion in that direction, and then cause a reverse motion. Thus an oscillatory motion would be produced, which in the case of no friction would never stop, but which in the case of friction is generally of short duration. If the surface of this fluid and all isobaric surfaces are perfectly horizontal, then *de*, Fig. 1, vanishes, and also the quotient e, and hence, also, the force ge at all depths. There are, therefore, no forces then to give rise to and maintain motions, and the fluid remains at rest.

63. If different parts of the fluid of a canal or reservoir have different densities, there can be no static equilibrium of

the forces arising from gravity with any arrangement of the fluid, as long as the differences of density remain, but the forces give rise to, and maintain, a system of counter currents, which, however, tend to equalize the differences of densities unless they are being continually reproduced from some cause. The differences of density are usually caused, either directly or indirectly, by differences of temperature. Since heat expands all fluids, if different parts have different temperatures they are expanded unequally; and as density is inversely as the volume of the same unit of mass, the more the fluid is expanded by heat, the lighter any given volume of it becomes.

Let us consider here the simple case only of a canal or trough of definite length and of uniform depth. If all parts have the same temperature, and consequently the same density, we have seen that if the surface is perfectly level, all the isobaric surfaces are level, and there is no tendency in any part of it to flow in any direction, and the whole fluid remains at rest. If, however, one end of the canal has a greater temperature than the other, with such a distribution, let us suppose, that there is a uniform temperature gradient from one end to the other, and has also the same temperatures at all depths of the same parts of the canal, then the fluid of the warmer end is expanded more than that of the other end, and its surface and each of the isobaric planes which were level before expansion are raised up at one end so as to become inclined planes, and the inclination is greater in proportion to the height of the plane above the bottom. There is, consequently, a tendency in the fluid to flow down the inclined planes toward the colder end; but this tendency is not the same at all depths, as in the previous case of a disturbance of level but not of density, but is greatest at the top and gradually decreases downward, being in proportion to the height above, and vanishing at, the bottom. The mere upward expansion of the fluid at the warmer end, before any horizontal motion ensues, does not affect the pressure at the bottom, and there is still a uniform pressure there from one end to the other, and consequently no force there to move the fluid in either direction. As the upper part

of the fluid, however, begins to flow, it fills up a little the other end and raises its level and increases the pressure there on the bottom, while the level at the warmer end is depressed a little and the pressure at the bottom decreased; so that there is now at the bottom and in the lower strata a gradient of pressure decreasing from the colder to the warmer end, and a tendency in the fluid there to flow in that direction, and thus to give rise to a counter current in the lower strata. After this has taken place the arrangement of the isobaric surfaces, represented by the nearly horizontal lines, and the motions of the fluid, are somewhat as represented in Fig. 2, according to which these surfaces decline from the warmer to the colder end of the canal in the upper strata of the fluid, and reversely below, and in consequence of which the fluid flows in counter directions above and below a given neutral intermediate plane

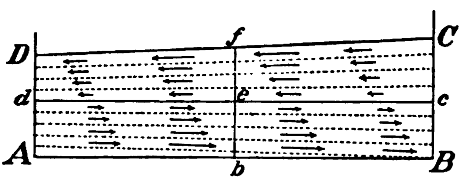

Fig. 2.

of equal pressure, *ced*, in which, being horizontal, there is no pressure gradient, and consequently no tendency in the fluid to move in either direction. In the upper strata the greatest velocity is at the surface, and this gradually becomes smaller with increase of depth below the surface until the neutral plane is reached, when the velocity vanishes and changes direction, and then increases with increase of depth below the surface; but in the case of friction between the fluid and the bottom of the canal, this velocity at a certain depth before the bottom is reached has its maximum, and then decreases again, and is least at the bottom.

64. In the motions of fluids the condition must be satisfied that no more of the fluid shall pass into any given space than flows out of it in a given time where there is no change of density, or more flow out than flows into it; for in the former

case the density would be necessarily increased, and in the latter diminished, or a partial vacuum produced. This is called *the condition of continuity*. In the counter motions of the canal this condition must be fulfilled, and this, consequently, after the regular motions are established and there is no further change of surface level, requires that just as much fluid in the counter currents must pass through any given vertical section *bef* of the canal in the one direction as in the other.

The satisfying of the condition of continuity not only requires counter horizontal motions in the canal, but likewise vertical ones; so that there must be a gradual settling down of the fluid toward the bottom at the end toward which the currents in the upper strata run, and a rising up again toward the surface at the other end. The vertical components of velocity, at least in a shallow canal, are extremely small in comparison with those of the horizontal motions, and are not indicated in Fig. 2. The counter vertical motions have likewise a neutral plane, a vertical one, *bef*, somewhere in the middle part of the canal, separating the part of the fluid which, at any time, has a slight downward tendency from that on the other side which is slowly rising toward the surface; and with reference to this plane the condition must likewise be satisfied that just as much of the fluid must slowly descend on the one side toward the bottom, through any given horizontal section of the canal as that represented by *de*, as rises up on the other side of it toward the surface, through the corresponding section *ce* at the same depth. If the canal or trough, therefore, were wedge-shaped, or had any shape which would make it narrower toward the one end than the other, the vertical section dividing the very slowly moving vertical counter currents, would have to be nearer the wide end so that the horizontal sectional areas of these counter vertical currents might be equal, or nearly so.

65. If there were no frictional resistances to the counter motions, the tendency of the pressure gradients, in the one direction above and the contrary below, would be to cause a continued acceleration of these motions; but as there are al-

ways resistances of this kind, the motions are only accelerated until the frictional resistances become equal to the forces causing and maintaining the motions, after which the whole of the forces is spent in overcoming the resistances, and there is no further acceleration.

In case of friction between the horizontal strata, but none between the fluid and the bottom of the canal, the velocities of the counter horizontal currents must increase from the horizontal neutral plane both toward the top and the bottom, where they are, consequently, the greatest. The relative velocities, however, between the strata must decrease from this neutral plane both upward and downward, and be least at the top and bottom. The forces arising from the pressure gradients which cause the counter motions act on all the strata, but most upon the upper and bottom strata, since here the gradients are greatest. The action of the force upon the first stratum increases the relative velocity between this stratum and the next until the frictional resistance is equal to the force. This force, by means of friction, is communicated to the second stratum, so that the force overcoming the frictional resistance between the second and third stratum is equal to the sum of the forces acting upon the first two strata, and consequently the relative velocity between the second and third is greater than that between the first and second. So the force overcoming the resistance between the third and fourth is equal to the sum of the forces acting upon the first three strata, and so on. The relative velocities between the strata must therefore increase from the surface down to the neutral plane where the forces vanish. But the sums of these forces are not proportional to the depth, since the gradients and the forces gradually decrease down to, and vanish at, the neutral plane. The relative velocities, therefore, at and near this neutral plane are greatest and vary very slowly near, and are least at, the top. The same would be the case commencing at the bottom, if there were no friction between the bottom of the canal and the lower stratum, so that this would be as free to move from the force acting upon it as the upper stratum, and

the actual velocities would be greatest and the relative velocities least at the bottom.

When the lower stratum suffers resistance from the bottom the velocities of the strata near the bottom increase up to some intermediate stratum between the bottom and the neutral plane, and then decrease up to this plane, where they vanish and become reversed in direction. In this case the resistances to the motions of the strata between the neutral plane and the bottom, if there is the same amount of flow through equal sectional areas, are much greater than they are in the strata above this plane, since in the one case the flow is resisted through friction both by the bottom of the canal and the neutral stratum, while in the other it is resisted by the one only, the neutral stratum. In order, therefore, that the condition of continuity may be satisfied in the case of friction on the bottom, there must be either a greater force, or a greater vertical sectional area for the fluid to pass through, or both, in the lower current; and both of these are brought about by an elevation of the horizontal neutral plane, that is, by a greater filling up and raising up of the surface of the colder end of the canal, which both makes the sectional area and the pressure gradients of the lower one of the counter currents greater in comparison with those of the upper one. There is then a slower motion through a greater sectional area in the lower current and a comparatively rapid one of smaller sectional area in the upper one, and the velocity is especially much greater near the surface.

66. The interchanging horizontal motions of the fluid between the two ends of the canal are oscillatory. The motion of any particle of the fluid starting from its extreme position in the warmer end and upper strata of the canal is first accelerated until the middle part is reached, after which it is gradually retarded until the particle is brought to rest in its extreme position toward the other end, where, gradually settling down toward the bottom, its motion is first accelerated again, and after passing its middle position it is then gradually retarded, and at the same time rises up into the upper strata whence it

started. Thus a particle describes a very elongated elliptic orbit, in which, especially in shallow and long canals, the transverse axis is very small in comparison with the other. The major axes also are very different for different particles, those near the top and bottom passing nearly or quite to the ends of the canal, while those near the central part and at medium depth make only short excursions on either side of the neutral vertical plane *bef*, Fig. 2. If, however, there are any abnormal disturbances, however small, besides that arising from the regular temperature gradient, or any irregularities in the bottom or sides of the canal, there is an interchanging and mixing up of the particles, so that the same one does not continue in the same regular orbit.

67. In the case of the open ocean extending from the equator to the pole, we can consider alone any part included between two meridians meeting at the pole, and this half-lune is similar to the wedge-shaped trough except that it does not become narrow so rapidly near the equator, and only becomes reduced to half its equatorial width on the parallel of 60°, which is two thirds of the distance from the equator to the pole. If we suppose that there is a uniform temperature gradient at all depths between the equator and the pole, the interchanging motion for any such half-lune would be somewhat as in the case of the wedge-shaped trough. There would be a more rapid motion in the upper current of smaller sectional area, and a comparatively slow one in the lower counter current with large sectional area; so that the neutral plane between the counter currents would be at no great depth below the surface. The vertical section, also, between the parts having a slow descent toward the bottom in the polar region and those slowly rising toward the surface in the equatorial region would not be on the parallel of 45°, but somewhere near the parallel of 30°, so as to make the horizontal sectional areas of both parts about equal.

DISTRIBUTION OF TEMPERATURE OVER THE EARTH'S SURFACE.

68. Since the motions of the atmosphere depend upon differences of temperature, it is necessary to know at the outset, in the investigation of these motions, the distribution of temperature over the earth's surface, not only for the mean of the year, but also for the different seasons of the year. Since all parts of the earth's surface on the same parallels of latitude have the same relation to the sun, upon which their temperatures and the differences of temperature depend, if the whole surface were either all land or all water, and perfectly homogeneous, there would be the same temperature all around the globe on the same latitude, and there would be, at all longitudes, the same temperature gradient between the equator and the pole. In the case of a wholly land surface, however, the difference of temperature between the equator and the pole, and consequently the temperature gradient, would be much greater than in the case of an earth entirely covered by the ocean. For in the latter case, in consequence of the difference of temperature of the water of the equatorial and the polar regions, there are interchanging counter currents, as just explained, by which heat is conveyed from the former to the latter region, the effect of which is to make the difference of temperature between the two regions much less than it otherwise would be, and consequently to diminish the temperature gradient between the equator and the poles. In the case of the earth, therefore, as really constituted, with continents and oceans extending from the equator to the pole, or nearly so, the temperature gradients between the equator and the pole on the continents are somewhat as they would be in case of a wholly land surface, while on the oceans they are somewhat as on an earth entirely covered by the ocean, and consequently the temperature gradients on the former are greater than on the latter. For this reason the equatorial and tropical parts of the continents are warmer than those of the ocean, while the reverse is true in the higher latitudes, and hence there are

differences of temperature on the same latitude in different longitudes.

In the general circulation of the atmosphere all the secondary and abnormal irregularities are neglected, and only the normal and principal temperature inequality, determined by the averages of temperature of each latitude, taken all around the globe, is considered. These averages of temperature are called *the normal temperatures of the latitude.* These normals of temperature, taken from "Meteorological Researches," Part I,[*] for the average of the year, and also for the two months of extreme temperatures, are given for each tenth degree of latitude in the following table:

Latitude.	January.		July.		Mean of the Year.	
	° C.	° F.	° C.	° F.	° C.	° F.
+80	−31.9	−25.4	+1.0	+33.8	−15.5	+4.1
70	26.5	15.7	6.9	44.4	9.8	14.4
60	16.9	−1.6	13.8	56.8	−1.6	29.1
50	−6.0	+21.2	18.6	65.5	+6.3	43.3
40	+4.5	40.1	22.8	73.0	13.6	56.5
30	12.9	55.2	26.6	79.9	19.8	67.6
20	21.7	71.1	29.0	84.2	25.3	77.5
+10	25.9	78.6	28.4	83.1	27.2	81.0
0	27.3	81.1	26.1	79.0	26.7	80.1
−10	27.9	82.2	24.0	75.2	25.9	78.6
20	26.6	79.9	20.8	69.4	23.7	74.7
30	23.0	73.4	15.6	60.1	19.3	66.7
40	17.6	63.7	11.1	52.0	14.4	57.9
50	11.1	52.0	+6.4	43.5	8.8	47.8
−60	+3.6	+38.5	0.0	+32.0	+1.8	+35.2

It is seen from this table that, while the temperatures at and near the equator are very nearly the same the year around, in the higher latitudes of the northern hemisphere they vary very much in the different seasons of the year, and that the temperature gradients between the equator and the pole are more than twice as great in midwinter as in midsummer. In the southern hemisphere, however, these annual variations are comparatively very small. This is because nearly all the land lies in the northern hemisphere and the southern hemisphere

is mostly covered by the ocean, which equalizes somewhat the extreme temperatures of the seasons.

On account of the interchange of equatorial and polar waters in nearly every part of the southern hemisphere, while this in the northern hemisphere takes place over a much smaller part, since much less of it is covered by the ocean, the normal temperatures of latitude for the average of the year in the southern hemisphere are a little less in the lower, and a little greater in the higher, latitudes than on the corresponding latitudes of the northern hemisphere. This causes the maximum of temperature or thermal equator to fall a little north of the equator.

The normals of latitude in the preceding table were obtained from Buchan's isothermal charts. According to later researches of Dr. Hann,[2] in which are used observations on high southern latitudes more recent than any used by Buchan, and the tabular means and normals of temperature given by Herr Spitaler, deduced from Dr. Hann's recent isothermal charts, the normals of temperature in these latitudes are somewhat less than those given in the preceding table. Although the disturbing forces which we are to consider in the motions of the atmosphere do not depend upon the absolute average temperature of the globe but upon differences of temperature or temperature gradients, yet it may not be amiss to state here, as a matter of general interest, that the average temperature of the atmosphere at the earth's surface, taken over the whole of the northern hemisphere, is 15°.3 (59°.5 F.), and that that of the southern hemisphere is sensibly the same.

69. It must not be understood that the preceding normals are those of the whole atmosphere extending to the top, but simply those of the part in contact with the earth's surface, for the temperature everywhere decreases with increase of altitude. The rate of this decrease, on the average for all latitudes and all seasons of the year, is, for the lower half of the atmosphere, about 0°.65 for each 100 meters of ascent; or, in English measures, 0°.36 for each 100 feet. This rate is supposed to be a little

greater in the lower than in the higher latitudes, but the difference is small, and it is also greater in general in summer than in winter, and by day than by night, and this is especially the case on land and in the lower part of the atmosphere. The variation of temperature with altitude, however, so long as the unstable state is not induced, is of little importance in the dynamics of the atmosphere, since the forces depend not upon absolute temperatures, but upon the horizontal temperature gradients, and these, so far as they pertain to the general circulation, are very nearly the same at all altitudes.

DISTRIBUTION OF AQUEOUS VAPOR.

70. Since aqueous vapor under the same pressure is lighter than air, its density being to that of air as 0.622 to 1, the unequal distribution of this vapor in the atmosphere also gives rise to small differences of normal pressure in different latitudes, or, in other words, to pressure gradients between the equatorial and polar regions, but these are small in comparison with those arising from difference of temperature. It is seen from Table II, Appendix, that the greatest vapor tension which can exist in the atmosphere at the equator with a temperature, say, of 26°, is about 25 mm.; while for the cold temperature of the polar regions at a temperature of $-15°$ the tension of saturation is almost nothing. As the relative humidity of the atmosphere at all places and all seasons of the year is about 80 per cent, this gives $25 \times 0.80 = 20$ mm. for the average vapor tension at the equator. This is the 1/38 part of 760 mm., the normal barometric pressure of the atmosphere. The density of the aqueous vapor being 0.622, the density of the part of the atmosphere comprising the aqueous vapor is diminished by $1 - 0.622 = 0.378$, and this multiplied into 1/38 gives a decrease of the density of the whole atmosphere equal to 1/100 part. This is equal to the diminution of density arising from an increase of temperature of 2°.7. Hence the effect of the average amount of aqueous vapor in the atmosphere at the equator on its density is about the same as

that of increasing its temperature 2°.7, or about the one-tenth part. This is the proportion of increase of the numerical coefficient of temperature which Laplace introduced into his barometric formula in order to take into account approximately the effect of the average amount of aqueous vapor in the atmosphere. Although this proportion is very nearly correct for equatorial temperatures at and near the earth's surface, yet it becomes erroneous for temperatures at and below the zero of the Centigrade scale; but at these temperatures the absolute amount of vapor is so small that the erroneous proportion gives rise to only very small errors. In all cases, therefore, we may assume, that for the average or normal state of the atmosphere the modifying effect of the aqueous vapor is such as to cause the atmosphere to expand its volume and diminish its density about the 1/250 instead of the 1/273 part, for each degree of increase of temperature. But this is the case in the lower part only of the atmosphere. In the upper regions of the atmosphere the proportion of vapor in the atmosphere is less, both on account of the diminished temperature with increase of altitude, and also because evaporation takes place at the earth's surface and the vapor is very slowly diffused upward to high altitudes. The decrease of density, therefore, from this cause, at high altitudes, and the corresponding effect upon the pressure gradients between the equator and the pole, are small in comparison with those arising directly from differences of temperature. Upon the whole, therefore, considering the whole depth of the atmosphere, the effect of the aqueous vapor is small, but even this depends indirectly upon difference of temperature, since without this there would be no difference in the amount of the vapor in the equatorial and the polar regions.

GENERAL CIRCULATION WITHOUT ROTATION OF THE EARTH.

71. In treating the general circulation of the atmosphere, it will be best to consider first, in a preliminary way, the more simple case of circulation without rotation of the earth on its

axis. For although this is not the real case of nature, yet the results obtained in this case will be useful and necessary in treating the more general and more complex case in which the earth turns on its axis. Let *PCE*, Fig. 3, represent a quadrant of a meridional section of the earth, and *aa'*, *bb'*, *cc'*, etc., represent infinitely thin strata of the atmosphere of equal pressure in case all parts had the temperature of the pole *P*. In this case these strata would all be horizontal or equidistant from the earth's surface and from one another at the equator and at the poles, neglecting quantities of the order of the earth's ellipticity, which are of no importance here. But by the preceding table,

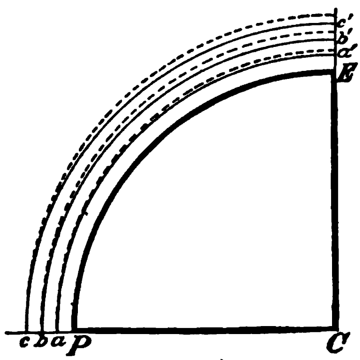

Fig. 3.

§ 68, the temperature of the atmosphere at the earth's surface is about 45° C. greater at the equator than at the poles; and in the upper strata, we have reason to think, the difference is nearly the same. The effect of increase of temperature being to expand the air the 1/273 part of its volume at the temperature of melting ice for each degree of its increase, the effect of an increase of temperature of 45° at the equator is to expand the air upward the 45/273 or 1/6 part nearly of its volume at this temperature, but a little more than this part of its volume at the temperature of the poles. The stratum *aa'* is accordingly elevated at the equator about 1/6 of the altitude *Ea'* of this stratum above the earth's surface, that of *bb'* by 1/6 of *Eb'*, and that of *cc'* by 1/6 of *Ec'*, and so on; each successive stratum

being raised in proportion to its altitude above the earth's surface, and the new positions of the strata, resulting from such an increase of temperature as given in the preceding table, and consequently the upward expansions, are somewhat as represented by the dotted lines in the figure.

If this increase of the temperature of the equatorial over that of the polar regions were made instantaneously, the first effect of the expansion would be to cause gradients down which the air of the upper strata would tend to flow from the equator toward the poles,—just as in the case of the canal, § 62, the upper strata have a tendency to flow from the warmer toward the colder end; and the greater the altitude above the bottom or earth's surface, the greater is this tendency in both cases. At the earth's surface, however, there would be no tendency in the air to flow either from or toward the equator, since the mere upward expansion would not affect the pressure at the earth's surface, and thus give rise to a pressure gradient acting in either direction.

The linear gradient e of any isobaric surface at the altitude h, due to difference of temperatures, is equal to the difference of upward expansion of a vertical column of air at each end of a unit of length in the direction in which the gradient is reckoned, whatever may be the assumed unit. Putting, therefore, τ_1 and τ_2 for the temperatures at the beginning and end of the unit respectively, we shall have the upward expansion of the former equal to $\frac{1}{273}h\tau_1$ and that of the latter equal to $\frac{1}{273}h\tau_2$, if h is the height of the column at the temperature of 0° C., since the volume of air expands the $\frac{1}{273}$ part of its volume at 0° C., for each degree of increase of temperature. Hence, putting $\varDelta\tau$ (variation of τ) for the change of temperature in the distance of unity, or, in other words, for this temperature gradient, we shall have

$$e = \frac{1}{273}h\varDelta\tau.$$

72. The first effect of the motion in the upper strata of the atmosphere from the equator toward the pole, as in the case of the water flowing from the warmer toward the colder

end of the canal, would be to fill up a little, as it were, the polar region with air from the equatorial, the effect of which is to increase the pressure a little in the former and to decrease it a little in the latter region, thus creating at the earth's surface and in the lower strata a gradient of pressure decreasing from the pole toward the equator, which would cause a counter current in the lower strata. The neutral plane where there is no pressure gradient, and consequently no motion in either direction, is now raised from the earth's surface up to such an altitude that the counter-flows between the equator and the pole, *from* the equator above and *toward* it below, satisfy the condition of continuity, and this altitude must be greater or less according to the relative amounts of friction in the upper and lower strata. For reasons given in § 65 in the case of the canal, this must be above half the mass of the atmosphere. Since the upper part of the atmosphere has no definite limit, but becomes more and more rare and extends to a very high but unknown altitude, the isobaric surfaces of the upper very rare part of the atmosphere have very steep gradients, which increase in proportion to the altitude above the neutral plane. But the force with which a given volume of air tends to slide down these gradients or press toward the pole is as the density, and this, where the temperature remains the same, is as the pressure. While the gradients, therefore, increase with increase of altitude above the neutral plane in an arithmetical progression, the forces for the same gradient decrease with increase of this altitude in a geometrical progression, and so become small above, while, by the nature of gases, § 33, the frictional resistances for equal relative velocities between the strata are the same for all densities, and so the same at all altitudes. Hence the relative velocities between the strata, in the case of no rotation of the earth on its axis, would be very small at great altitudes, and the absolute velocities nearly the same at all high altitudes.

73. The whole vertical circulation of the atmosphere in this case would be similar to that of the water of the canal, § 66, the particles of air would have an oscillatory horizontal and

vertical motion, which in the upper strata of the equatorial region is first accelerated until the particle arrives at some intermediate latitude, after which it is gradually retarded in its further motion toward the pole and brought to rest at its polar limit of range, meanwhile gradually settling down toward the earth's surface, where the pressure gradient is reversed and the reversed motion of the particle is again accelerated until it arrives at some intermediate latitude of its range of oscillation, after which it is again retarded in its horizontal motion and brought to rest, meanwhile rising up to higher altitudes, where the force of the pressure gradient is again reversed, and it commences again another circuit as before. The nearer the particle is to the neutral planes, the shorter are the ranges of oscillation, both horizontal and vertical, and the smaller the elliptic circuit. On account of the narrowing of the spaces between the meridians in approaching the pole, the vertical plane between the descending air in the higher latitudes and ascending air in the lower ones, as in the case of the oceanic circulation, cannot be half-way from the equator to the pole, but must be about the parallel of 30°.

In the case of no friction the motions would be continually accelerated as long as any difference of temperature between the equator and the pole remained, but of course a very rapid and increasing interchange would tend to diminish and finally to sensibly destroy this difference. In the case of friction the motions would be accelerated until the frictional resistances would be equal to the forces, after which a uniform horizontal oscillatory motion would be maintained, the relation between the gradients and the frictional resistances being such as to leave a small residual force to alternately accelerate and retard the motions of the oscillations.

GENERAL CIRCULATION WITH ROTATION OF THE EARTH ON ITS AXIS.

74. The general vertical circulation of the atmosphere which would take place in case the earth had no rotation on its axis having just been explained, we come now to consider the

effect upon this circulation of the deflecting force arising from the earth's rotation, explained in the preceding chapter, in the real case of nature. In consequence of this force, § 53, in whatever direction any part of the atmosphere may be moving, it is continually deflected, if free, to the right of the direction of motion in the northern hemisphere and the contrary in the southern; and if not free, it presses in these directions and causes atmospheric gradients, ascending in the directions of pressure. Hence the vertical circulation due to the real forces arising from difference of temperature between the equator and the poles being once established, the air in the upper strata of the atmosphere in either hemisphere, in moving toward the pole, is deflected eastward, and in the counter current in the lower strata the tendency is to counteract and destroy motion toward the east and then to cause a west component of motion. These contrary forces, acting toward the east above and the west below, give rise to relative velocities between the strata, and maintain them by overcoming the friction between the strata, which otherwise would constantly tend to reduce and to destroy these relative velocities, and to reduce the absolute velocities above and below all to the same velocity on any given parallel of latitude.

In the case of no friction between the air and the earth's surface, the east and west components of velocity arising from the interchanging motions between the equatorial and polar regions and the deflecting forces depending upon the earth's rotation arising from them, would satisfy the expression

$$rw = c,$$

in which the value of c is the average value of rw for all the particles of the air of the hemisphere, before interchanging motion took place, the definitions of r and w being those of § 47; or it is the average value of $r_0 w_0$, in which r_0 and w_0 are the values of r and w before motion commences and while the air is at rest relative to the earth's surface, in which case w is equal to rn. If each particle of air were entirely free, both from the effect of friction of the earth's surface and the inter-

action of the particles upon each other, then its motions would satisfy the equation $rw = r_0 w_0$, and since in the interactions between the particles action and reaction are equal, these cannot affect the average value of rw for the whole hemisphere, and it must remain the same at all times and be the same as before motion commenced, that is, be the average value of $r_0 w_0$ taken for the whole hemisphere.

In order to satisfy the preceding relation near the pole, where r is small w must be very large; and this here becomes mostly an east velocity relative to the earth's surface, since near the pole the part depending upon the earth's rotation, rn, is small.

In consequence of friction between the lower stratum of the atmosphere and the earth's surface, this stratum, with the forces only being here considered, cannot have either an easterly or a westerly motion, since the rate of polar motion on any given parallel, mass multiplied into velocity, in the upper strata in which the air moves from the equator toward the poles, is, on account of the condition of continuity, exactly equal to that of equatorial motion in the lower strata, in which the air moves from the poles toward the equator; and so the whole force deflecting eastward above, arising from the earth's rotation, is exactly equal to that deflecting westward below, and these being in contrary directions, there is no force arising directly from the interchanging motions and deflecting forces to overcome the friction between the atmosphere and the earth's surface. This force arises from the vertical motions, as explained further on in § 77. But whatever east or west components of velocity may exist at the earth's surface, there are the same relative velocities of these components all the way up above, so that all the absolute east or west components above are increased algebraically by the amount of velocity at the surface.

75. The greater the absolute east components of velocity, the greater also are the relative velocities and the friction between the strata, so that for a given amount of interchanging motion between the equatorial and polar regions there is a

limit beyond which these absolute east components of velocity cannot go, and this limit is where the frictional resistance between the strata becomes equal to the forces which overcome it and maintain the motions. But now with the establishment of these east components of motion, increasing with increase of altitude, there is called into play another deflecting force depending upon the earth's rotation, which interferes with and diminishes the amount of interchanging motion between the equatorial and the polar regions which would exist if the earth had no rotation on its axis. For as in the case of a current of air flowing from the equator toward the pole in the northern hemisphere there is a force deflecting to the right or east, so in the case of the east component of motion there is also a force deflecting toward the right or the equator. In the southern hemisphere it is to the left, but still toward the equator. This force is contrary to, and counteracts, that arising, as explained in § 71, from the temperature gradient between the equator and the pole. The greater the east components of velocity in the atmosphere above, the greater this force; so that if certain velocities were reached, this force would be exactly equal to that depending upon the temperature gradient, and the interchanging motion between the equator and the poles could not then be maintained, and without some motion of this sort there would be no forces to overcome the friction and maintain the east components of motion. The limit, therefore, beyond which the east components of velocity cannot go must fall a little short of those velocities which would entirely destroy the force which keeps up the interchanging motions; but the less the frictional resistance to the motions the more nearly they can approach to this limit, and in the case of no friction the east components of velocity which would satisfy all the conditions of the problem would be those coming up to the limit; for in this case, the interchanging motions being once established, there would be no need of any force to overcome friction and to maintain them.

The general motions of the atmosphere and their relations to the forces producing them, in the case of an earth with ro-

tation on its axis, are similar to those of machinery run by a steam-engine, and controlled by a governor. Without this the whole force would be applied to overcome the friction and doing the work, and great speed might be obtained, but with the governor suitably adjusted there is a certain limit beyond which the speed cannot go, however much the friction or work of any kind may be diminished; for when this is reached all the force is cut off. As long, therefore, as there is any friction to be overcome or work of any kind to be done, the speed must fall a little short of this limit, and it can only reach this limit when these vanish. So, in the general motions of the atmosphere in case of no rotation of the earth on its axis, we have seen, § 73, the whole force of the temperature gradient is brought to bear in overcoming the friction and maintaining the interchanging motion between the equatorial and polar regions, and so this might become very great. But in the case of the earth with rotation on its axis, there are soon developed east components of motion, which, by means of the consequent deflecting forces, in a great measure counteract or cut off the forces which give rise to and maintain the interchanging motion, so that there is a limit beyond which the speed of this motion cannot go; for if this limit were reached, the motions would give rise to east components of motion and deflecting forces which would entirely counteract the forces arising from the temperature gradients.

76. The limit beyond which the east components of velocity cannot go may be determined by equating the expression of the horizontal force for unit of mass, ge, § 62, with that of F_v, § 49, which is the deflecting force acting in the direction from the poles toward the equator, depending upon the east components of velocity v. We thus get

$$ge = (2n + \nu)v \sin l.$$

By neglecting ν in comparison with $2n$, as we may without any sensible error in the case of any real easterly motions in the general circulation of the atmosphere, we get from the preceding equation

$$v = \frac{ge}{2n \sin l}.$$

The same is deducible from the expression of e, § 56, by changing s to v, since this is simply the special case in the general expression in which s becomes v. This limiting value of v can be expressed in a function of the temperature gradient $\Delta\tau$, instead of that of e, the gradient of the isobaric surface, by putting for e above its value in § 71. We thus get

$$v = \frac{gh\Delta\tau}{273 \times 2n \sin l} = 246.4 \frac{h\Delta\tau}{\sin l},$$

using for g and $2n$ their numerical values found in § 45, in obtaining the last form of expression.

Strictly, the value of h in this expression is the height of the isobaric surface if the atmosphere had the temperature of 0° C.; but as the average temperature up to a considerable altitude, and for all latitudes, does not differ much from this temperature, this value of h in any case differs but little from the actual height of this surface, and so the latter may be used for approximate results without much error.

If $\Delta\tau$ be assumed to be the change of temperature in one degree of latitude, instead of one meter, then we must divide this expression by 111,111, the number of meters in a degree, in order to obtain the expression of v in terms of the temperature gradient thus expressed. We thus get

$$v = 0.00221 \frac{h\Delta\tau}{\sin l}.$$

Since $\sin l$ becomes small towards, and vanishes at, the equator, and $\Delta\tau$ is supposed to do the same, the value of v becomes indeterminate at, and very uncertain near, the equator, since, where $\sin l$ is very small, a small inaccuracy in $\Delta\tau$ would give rise to a large error in v.*

* If the temperature τ for each isobaric surface is a function of the latitude of the form $A_0 + A_1 \cos 2l$, in which A_0 is the temperature on the parallel of 45° where $\cos 2l = 0$, and A_1 is the half difference of the temperatures at the

77. So far we have considered only the horizontal motions and their deflecting forces in the vertical circulation arising from difference of temperature between the equator and the poles, and have obtained the preceding results.

We come now to consider the vertical motions, or components of motion, downward from the upper strata toward the earth's surface in the higher latitudes, and the reverse in the lower latitudes. As the air in the upper strata of the higher latitudes with its large east component of velocity above settles down toward the earth's surface where this component is less, it gradually loses its momentum, and the effect of this lost momentum transferred by means of friction from one stratum to another until it reaches the earth's surface, is spent in overcoming the friction between the lower stratum and the earth's surface and maintaining a small east component of velocity of the air at and near the surface.

In the passage of the air of the upper strata from the equatorial toward the polar regions, a part of the force deflecting eastward is spent in overcoming the frictional resistance between the strata with east components of velocity increasing with increase of altitude, which, we have seen, is necessary to satisfy the conditions, and the other part is spent in accumulating the momentum of the east components of motion. In passing in

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equator and the pole, then, by taking the differential of this expression and substituting the small variations of temperature and of latitude $\Delta\tau$ and Δl, respectively, for the corresponding differentials $d\tau$ and dl, we have, since Δl has been assumed to be equal to unity or one meter,

$$\Delta\tau = 2A_1 \sin 2l = 4A_1 \sin l \cos l.$$

Substituting this expression of $\Delta\tau$ in the first of the preceding expressions of v, we get

$$v = 985.6hA_1 \cos l.$$

Since v here is proportional to $\cos l$, and this to the distance of the air with velocity v from the axis of rotation, the velocities which, at any given altitude h, correspond to the limit of velocities which can be reached, are those which have the same angular velocity with reference to the earth's axis.

The values of A_1 corresponding to the mean of the year, and January and July respectively, according to the table of § 68, are approximately 22°, 30°, and 14°, as deduced from the northern hemisphere. The temperature gradient, therefore, is more than twice as great in January as in July.

the lower strata from the polar toward the equatorial regions, the reverse is the case, and part of the force now tending to deflect westward is spent in overcoming the frictional resistances to a west component of motion, whether arising from the strata above having greater east components of velocity, or from the earth's surface below, where, in the lower latitudes, there is a west component of motion, and the balance is spent in overcoming the momentum of the east component of motion in the higher latitudes already acquired, or in giving rise in the lower latitudes to the momentum of the west components of motion. In the higher latitudes, therefore, where there is an east component of motion at the earth's surface, the amount of east momentum lost by any portion of air in its passage above from any given latitude toward the pole until it returns below to the same parallel, is the kinetic energy which it has contributed toward overcoming the frictional resistance of the earth's surface to the east component of motion at the surface between that parallel and the pole; and in the lower latitudes, where there is a west component of motion at the earth's surface, the amount of west momentum lost or east momentum gained, or, considered algebraically, the amount of west momentum lost, by any portion of air from the time it leaves any given parallel of latitude in its passage toward the equator below until it returns to the same parallel above, is the amount of kinetic energy which it has contributed towards overcoming the frictional resistance of the earth's surface to the west components of motion between that parallel and the equator.

From the consideration, therefore, of the interchanging horizontal and vertical motions and the change of east and west momentum as the vertical motions take place, we see there is a sort of torsional force arising from the effect of the earth's rotation on the interchanging motions arising from the temperature gradient, which tends to give rise to and maintain an east component of motion of the atmosphere at the earth's surface in the higher latitudes, and the reverse in the lower latitudes. There are, however, the same relative velocities between the successive strata, since the forces remain the

same which overcome the friction and keep up these relative velocities, as if there were no such torsional force, or the frictional resistance to east or west components of motion at the earth's surface were infinitely great. Whatever velocities, therefore, of this sort are given to the lower stratum of the atmosphere, the same increase of velocity is also given to all the strata up to the top of the atmosphere. If we therefore put v_0 for the value of v at the earth's surface, instead of the expression of v_0 in § 76, we now get

$$v = v_0 + 246.4 \frac{h\Delta\tau}{\sin l},$$

in which $\Delta\tau$ is the change of temperature in the distance of one meter. If $\Delta\tau$ expresses the change in the distance of one degree of latitude, then 0.00221 must be used instead of the numerical coefficient 246.4.

According to this expression of v, the east components of velocity in the higher latitudes where v_0 is positive increase from the earth's surface to the top of the atmosphere. In the lower latitudes where v_0 is negative or a west component the west components of velocity above the earth's surface decrease up to a certain altitude, where they vanish and change to east components, and then these increase to the top of the atmosphere. It is evident that the higher the stratum the nearer to the equator must be the parallel of latitude where the change from west to east components of motion takes place until that altitude is reached where there is no west component of motion at any parallel.

78. It is seen from this expression that the differences between the limiting east components of velocity at different altitudes, and approximately, at least in the higher latitudes, between the real velocities, are proportional to $\Delta\tau$, and so to the difference of temperature between the equator and the poles. But the amount of east or positive momentum lost in a given time while the air in the higher latitudes is descending from higher to lower altitudes, and west or negative momentum, in ascending in the lower latitudes from lower to higher

altitudes, depend both upon the rate of change in the east components of velocity with regard to change of altitude and also upon the speed of the vertical circulation. This latter, if we suppose the friction between the strata, moving with relative velocities, to be as these velocities, is, as likewise the former, proportional to $\varDelta\tau$, the temperature gradient which gives rise to the forces overcoming the friction and maintaining the circulation. The amount of east or west momentum lost in a given time, and spent in overcoming friction between the atmosphere and the earth's surface and giving value to v_0 is, therefore, as the square of $\varDelta\tau$, or as the square of the difference of temperature between the equator and the poles. If, then, frictional and other resistance between the atmosphere and the earth's surface were as the velocity, the value of v_0 would have to be as the square of $\varDelta\tau$, and so about four times greater in winter than in summer. But it is probable that the resistances at the earth's surface increase in a ratio which is greater than the first power of the velocity, and if so the value of v_0 would have to increase in a ratio which is less than that of the square of $\varDelta\tau$. But whatever may be the relation between the frictional and other resistances and velocity, the values of v_0 must be greater in winter than in summer.

The relative east velocities between the motions of the upper strata and the earth's surface are likewise greater in winter than in summer, on account of the value of $\varDelta\tau$ in the expression of v in § 77 being greater, and in the northern hemisphere more than twice greater in January than in July. There is, then, an annual inequality in the east and west components of motion of the atmosphere, both at the earth's surface and all altitudes above the surface, such that the east components of velocity above are greater in January than in July, and in the middle latitudes of the northern hemisphere, more than twice as great, since the easterly velocities there at the earth's surface are very small.

In the case of a stratum of air between two parallel and vertical plates extending from the equator to the poles and reaching to the top of the atmosphere, the horizontal inter-

changing motion between the equatorial and polar part of it arising from a vertical circulation would not give rise to any force which would tend to turn the polar end toward the east and the equatorial toward the west, if it were free to turn around some vertical axis in its middle. For in this case there would be no east momentum lost in the descent of the air in the polar end, or west momentum, in its ascent at the equatorial end, and the deflecting force of the current of air moving toward the pole above, and causing pressure against the east side, at any latitude, would be exactly counteracted by that of an equal amount of motion toward the equator below, and causing pressure against the west side. In this case no part of the deflecting forces is spent in creating east momentum in lower latitudes which is lost in higher latitudes, and *vice versa,* but it is all spent in giving rise to lateral pressure.

79. There is still another effect in addition to the preceding, though much smaller, depending upon the ascending and descending components of motion in the vertical circulation, which tends to give rise to and maintain an east component of motion in the higher latitudes, and the contrary in the lower ones. This is a necessary consequence of the principle of the preservation of areas in the case of central and centrifugal forces. As the air in the higher latitudes descends toward the earth's surface it comes nearer to the earth's axis; and so, if free, must have a proportionally greater absolute gyratory velocity around this axis, and consequently a greater relative velocity with reference to the earth's surface.

If not entirely free, this tendency to acquire an increased east component of velocity causes pressure in that direction, which tends to overcome the resistances to motion, whether frictional or otherwise, and through friction between the different strata is communicated to the earth's surface. In the lower latitudes where the air ascends the effect is the reverse, and tends to overcome frictional resistances to motions from east to west, and is likewise communicated through friction to the earth's surface. In the preceding case of a vertical stratum of air, free to turn around a vertical axis, these forces being in an

east direction in the higher latitudes where the air descends, and the contrary in the lower latitudes where it ascends, would both tend to turn the vertical stratum horizontally around the vertical axis. The formula for computing the absolute gyratory velocity of a projectile or any free body, forced directly upward, and consequently the loss of absolute velocity, and the relative west velocity acquired at any given altitude, is given in § 60. By reversing this formula, the amount of relative east velocity acquired in falling through a given space, if the body is entirely free, can be likewise computed. On account, however, of the small depth in comparison with the earth's radius of that part of the atmosphere having any considerable density, these effects are very small for the atmosphere generally.

80. The relation between the amount of the east components of velocity of the air at the earth's surface in the higher latitudes, and that of the west components in the lower latitudes, is determined by the condition that the sum of all the forces of the air-particles of the higher latitudes acting in an easterly direction upon the earth's surface through friction or resistance of any kind, multiplied into their distances from the axis of rotation, called moments of couple, must be exactly equal to similar products in the lower latitudes, where there is a west component of motion, and the forces by means of frictional and other resistances act in the contrary direction. If this condition were not satisfied there would be a residual uncounteracted force acting in the one direction or the other tending to change the velocity of rotation; but this can only arise from external forces which do not act in the direction of the axis of rotation, and not from any central forces, or such as act only in the planes of the meridians, and which consequently have no gyratory components of force, for the actions and reactions in the motions arising from such forces, being exactly equal and in contrary directions, where the mass as a whole remains at the same distance from the centre, cannot affect the velocity of rotation. This principle was recognized by Hadley in his theory of the trade-winds, for he states that "all motions in any direction must have their counter-motions,

else the effect upon the earth's surface would be to change the earth's rotation upon its axis."

The forces exerted upon the earth's surface tending to turn the earth on its axis are effective in proportion to the distance from the axis at which these forces are applied. This is not only in accordance with a well-established principle in mechanics, but also with almost every one's experience. One man at the end of a lever of a given length can turn a capstan which would require two men of equal strength with levers of only half the length. The forces, therefore, have to be multiplied into the distances from the axis, and then the forces of equal products have equal tendencies to turn the earth on its axis. If a given gyratory force were applied on the parallel of 60°, where the distance from the axis is only half as great as at the equator, its effectiveness in turning the earth on its axis would be counteracted by half the amount of this force applied in the contrary way at the equator, and it would require twice as much of the former to counteract the latter. The east components of velocity, therefore, in the higher latitudes, where the distances from the axis are much less, must be much greater than the west components in the lower latitudes where, the distances from the axis are much greater, if we suppose the surface of the earth to be homogeneous, so that equal velocities meet with equal resistances and so exert equal forces upon the earth's surface, or else the east components of velocity must comprise a much greater amount of air. The quantity of air is equally divided by the parallel of 30°, and therefore the east components of the higher latitudes at the earth's surface must be much greater than the west components in the lower latitudes, or else the dividing parallel between the east and west components of motion at the earth's surface must be nearer the equator, on the average, than the parallel of 30°.

81. The velocities of the east and west components at the earth's surface depending upon the forces of momentum, as explained in § 77, depend very much upon the nature of that surface. For the velocities increase until the resistances to the motions are equal to the forces producing them, so that

the smoother the surface, and the less the friction between the air and that surface, the greater the velocities. If there were no resistances from the earth's surface these velocities would be very great—somewhat as given in § 48, but on account of the smallness of the forces giving rise to them, and maintaining them by overcoming the friction when they are once established, these velocities are comparatively very small. Since they depend upon the amount of resistance to the motion, they must, in general, be greater on land than on the ocean on the same latitudes, and since there is little land in the southern hemisphere in comparison with that of the northern, they must be much greater in the former than in the latter. The values of v_0, therefore, in the expression of the limiting value of v, § 77, are much greater on the several parallels of latitude of the southern hemisphere than on corresponding parallels of the northern hemisphere, especially for the higher latitudes, where, in the southern hemisphere, the ocean extends entirely around the globe. The east components of velocity, therefore, at all altitudes in the higher latitudes, where v_0 is positive, or at least the limits of these velocities, must be greater in the southern than in the northern hemisphere by the same amount that v_0 is, since the relations between the east components of velocity of the strata above the earth's surface remain the same whatever those velocities may be at the surface.

The whole system of atmospheric circulation is so complex and the friction constant of the atmosphere and the resistances from the earth's surface so uncertain, that it is impossible to compute the values of v_0 from the known forces; and all that can be obtained from theory is an indication that there must be motions in certain directions, and that these must be greater or less according to the variations of the forces at different times and the resistances of the earth's surface at different places. Even the values of v_0 are very imperfectly known from direct observation, though this indicates that there are such values, positive in the higher and negative in the lower latitudes.

82. We have seen that in the case of no frictional resist-

ances to the motions of the atmosphere the east components of velocity of the different strata on any given parallel of latitude come up to the limit as given by the expression of v in § 77, whatever the surface value v_0 may be, and that in this case there is no necessity for any meridional interchanging motion to overcome the friction of these east components of motion. In the case of friction, therefore, the less the friction, and the greater the deflecting force belonging to a given amount of meridional motion, the less of this motion is required to overcome the friction of the east component of motion. The frictional resistance to the motions of the atmosphere, except near the earth's surface, are so small that a very little interchanging motion between the equatorial and polar regions is required to give force enough to overcome the resistance to the east component of motion, especially in the higher latitudes where sin l in the expression of this force, § 49, is large. In the higher altitudes, therefore, where the east components of velocity are great in comparison with the velocity of the interchanging motion, the resultant direction must be very nearly from west to east, being a little north of east in high altitudes where the interchanging motion is *toward* the pole, and a little south of east in the lower strata where it is *from* the pole.

In the lower latitudes, however, where the deflecting forces depending upon the earth's rotation are small in consequence of the smallness of sin l, the relative velocities between the strata, and the absolute east components of velocity in the strata above, are less than in higher latitudes for the same interchanging velocity, and so the resultant motion deviates more from an east direction. The motions here, both in velocity and direction, become more nearly such as they would be in the case of an earth without rotation, in which case they would be in the planes of the meridians and would become comparatively very rapid. At the altitudes in these latitudes where the west component of motion changes to an east component there is very little motion in any direction, and so the direction is very uncertain; but at still lower altitudes, where there is a west com-

ponent of motion, this combined with the component of motion toward the equator gives rise to the trade-winds in the lower strata of these latitudes.

At the equator the deflecting forces giving rise to east motions entirely vanish, and hence here there are no such motions; and as the horizontal interchanging motions also vanish here, or at least at the thermal equator, on account of the vanishing of the temperature gradient, there can be no such motion here. Though there may be a westerly motion at a considerable altitude above the earth's surface arising from the unchecked momentum of the west component of motion acquired as the air moves in from both sides toward the equator.

COMPARISONS WITH OBSERVATIONS.

83. The preceding theoretical deductions, so far made, with regard to the general circulation of the atmosphere, must be understood to be those simply which depend upon the normal temperature gradient between the equator and the poles, unaffected by abnormal local and temporary disturbances of temperature. Hence in these comparisons the prevailing motions and directions of the air, observed at different times and places, and not individual observations, made in one locality, must be used.

With regard to the strong easterly currents of the upper strata of the atmosphere in nearly all latitudes, as deduced from theoretical considerations in the preceding pages, any one of ordinary observing habits could scarcely live a week upon the earth without discovering from the motions of the clouds, and especially the very high cirrus clouds, that the general tendency of the air above is easterly, even in the lower latitudes where, in the lower strata, there is a west component of motion, and that the velocities above in the middle and higher latitudes are much greater than at and near the earth's surface. Espy says: "I have found the true cirrus clouds to average scarcely once a year from any eastern direction; and when they

do come from that direction, it is only when there is a storm of uncommon violence in the east." Mr. Ley also, in his numerous observations of the cirrus clouds, almost universally found them to have an easterly motion, from which they rarely deviated. From these observations it is evident that the prevailing normal currents in the regions of these clouds is easterly, and with a velocity so great that abnormal temporary disturbances of the atmosphere are rarely sufficient to reverse their directions.

At Toronto, Canada, the resultant directions of the clouds, as ascertained from a very long series of observations, are as follows:

Spring.	Summer.	Autumn.	Winter.
N. 83° W.	N. 75° W.	N. 81° W.	N. 78° W.

Here, at all seasons of the year, the resultant directions of cloud-motion are from a direction a little north of west, indicating that the principal component of motion is an east component in accordance with theory. The other component, being a southern one, indicates that the clouds observed were mostly down in the lower strata of the atmosphere, below the neutral plane, where the interchanging motion is from the pole toward the equator.

That there is a strong easterly current prevailing at great altitudes in the atmosphere, deviating but little from an east direction, even as near the equator as the parallel of 30°, which the great local and temporary disturbances to which the atmosphere is subject scarcely ever reverses, or even changes much in direction, is shown from the results of four years' observations of the directions of the cirrus clouds at Zi-ka-wei,[11] latitude 31° 12′ N., longitude 121° 26′ E. These give, for the four years,

Directions.............	N.	N.N.E.	N.E.	E.N.E.	E.	E.S.E.	S.E.	S S.E.
Frequency.............	13	1	6	1	5	0	2	1
Directions.............	S.	S.S.W.	S.W.	W.S.W.	W.	W.N.W.	N.W.	N.N.W.
Frequency.............	3	6	43	69	395	24	18	1

Here the direction of by far the greatest frequency is the

western one, comparatively few being even from the adjacent directions given; and as there are a few more motions from W.S.W. than from W.N.W., and from S.W. than from N.W., they indicate that the normal current has a small northern component. The results, therefore, are strictly in accordance with theory, § 82, which requires that there should be a motion toward the pole at high altitudes, with a velocity small in comparison with that of the east component of motion.

Even much nearer the equator, at Colonia Tover, Venezuela, latitude 10° 26′, observations of the directions of the clouds indicate that the principal component of motion above is an eastern one, while in the lower strata the motions have a west component, as required by theory. While the motions of the upper clouds were generally from some westerly point, those of the lower ones were from some easterly direction."[8]

84. The theoretical deductions of the preceding pages are also confirmed by the observations of the directions in which the smoke and ashes of active volcanoes have been carried. On the first of May, 1812, the island of Barbadoes was suddenly obscured by a shower of ashes from an eruptive volcano of St. Vincent, West Indies, more than a hundred miles to the westward."[3] Although the motion of the air here in the lower strata has a western component, yet the ashes were carried up to altitudes where the principal component of motion is toward the east, and hence the ashes were drifted in that direction. Also on the 20th of January, 1835, the volcano of Coseguina, Central America, lying in the belt of the north-easterly trade-winds, sent forth great quantities of lava and ashes, and the latter were borne in a direction just contrary to that of the surface winds, and lodged on the island of Jamaica, 800 miles to the east-northeast."[3] In this case, also, the ashes were carried up to altitudes where the prevailing direction of the current of air was a little north of east, the northern component arising from the general poleward motion of the air at all latitudes in the upper strata of the atmosphere.

With regard to the volcanic eruption of the island of

Sumbawa, about 200 miles east of the eastern part of Java, Lyell says:[14]

"On the side of Java the ashes were carried to the distance of 300, and 217 miles toward Celebes, in sufficient quantities to darken the air. The floating cinders to the westward of Sumatra formed on the 12th of April a mass two feet thick and several miles in extent, through which ships with difficulty forced their way.

"The darkness occasioned in day-time by the ashes in Java was so profound, that nothing equal to it was ever witnessed in the darkest night. Although the volcanic dust when it fell was an impalpable powder, it was of considerable weight when compressed: a pint of it weighed twelve ounces and three quarters. 'Some of the finest particles,' says Mr. Crawford, 'were transported to the islands of Amboyna and Banda, which last is about 800 miles east from the site of the volcano, although the southeast monsoon was then at its height.' They must have been projected, therefore, into the upper regions of the atmosphere, where a counter current prevailed."

Notwithstanding this volcano was only a few degrees from the equator, yet it seems the finer ashes were carried to so great a distance eastward by the currents above, while the southeast trade-wind below, strengthened by the southeast monsoon prevailing at the time, carried the coarser particles in nearly the contrary direction.

In the account of the volcanic eruptions of Krakatoa of the latter part of May, 1883, it is stated:[15]

"The volcano was ejecting, with a great noise, masses of pumice, molten stone, and volumes of steam and smoke, part of which was carried away westward by the monsoon wind, dropping all around and close at hand its larger pieces, while a high rising cloud is especially recorded as driving away eastward, having evidently encountered a current in that direction in the upper air. Some of this dust-cloud was carried far to the eastward, for Mr. Forbes relates that on the morning of the 24th of May, when in the island of Timor, twelve hundred miles distant, he observed on the veranda of his hut, situated high in the hills behind Dilly, a sprinkling of small particles of grayish cinder, to which his attention was more particularly drawn later on, and the next day by their repeated falling on the page before him."

Here again, only a few degrees south of the equator, we

have evidence that while in the lower strata there was a current which carried the coarser parts of the ejecta westward, there was a strong upper current which carried the cloud of steam, smoke, and finer ashes, which ascended to great altitudes, away to the east.

85. The general easterly tendency of nearly the whole upper part of the atmosphere is not only shown from observed cloud-motions and the drift of the smoke and ashes of volcanoes, but also from observations of the winds on all high peaks and mountain ranges. Very strong and prevailing westerly winds have been observed on Mauna Loa in the Sandwich Islands, on the tops of Pike's Peak and of Mount Washington, in all the passes and on all the peaks of the Rocky Mountains and the Andes, on the Peak of Teneriffe, the Himalayas of Asia, and at every elevated position in either hemisphere all around the globe, except at and near the equator, even in the trade-wind zones, where, at the same time, there is a wind of considerable strength from nearly the opposite direction. Leopold von Buch says with regard to the Peak of Teneriffe:[15] "It is hard to find any account of an ascent of the Peak in which the strong west wind which has been met with on the summit has not been mentioned. Humboldt ascended the Peak on the 21st of June; when he reached the edge of the crater he could scarcely keep his feet, such was the violence of the west wind."

The following are the relative frequencies of the winds on Pike's Peak from the several directions, as given by 10 years of observations of the Signal Service 1873–1883, inclusive:[16]

Directions	N.	N.E.	E.	S.E.	S.	S.W.	W.	N.W.
Rel. Frequencies	8	13	2	2	4	36	17	18

From these results from the observations of direction merely, it is evident that at the top of the Peak the current, on the average, is easterly, with a considerable polar component. The latter is indicated by the relative frequency from the S.W. being twice as great as that from the N.W., though the contrary is indicated by the much greater relative frequency from the N.E. than from the S.E., but this should have much less weight. We

cannot infer, however, from the results of these observations that Pike's Peak reaches up to where the general motion is toward the pole, since the winds of this locality may be affected by an abnormal local disturbance of some kind.

For the top of Mt. Washington, from three years' observations, Loomis found the resultant direction of the wind N. 76° W. This and the resultants from cloud observations at Toronto, § 83, indicate that the general current at the top of Mt. Washington, and at the average altitudes of the clouds generally, has a south component of motion at these latitudes, as required by theory.

In the high latitudes of Asia it also appears that the direction of the wind on high mountain tops is easterly, and with a south component of motion. "At Mt. Alibut, 200 miles west of Irkutsk, and over 7000 feet high, a very constant and strong W.N.W. wind is observed." [20]

86. While theory indicates that the whole upper part of the atmosphere, except at and near the equator, must have an easterly motion with velocity increasing with increase of altitude, and the existence of such a motion is confirmed by all observations, it is evident from the Report of the Krakatoa committee of the Royal Society that at very high altitudes at and near the equator there must have been, in the latter part of August, 1883, a strong current from east to west of at least 70 miles per hour, by which the dust-particles causing the sunset glows and other phenomena were carried in this direction around the globe. According to the observations, also, of the Hon. Ralph Abercromby in the equatorial regions in the months of December and May, there is likewise a westerly current here at the altitudes of the cirrus clouds, though only a few degrees south of the equator there seemed to be an east component of motion. For such currents theory furnishes a slight force, § 60, arising from the ascent of air in the equatorial region, which tends to give rise to a westerly current, which, if it were not for friction, would be considerable at great altitudes. This current, it is seen, if the motions were without friction, as in the case of a projectile in a vacuum, would have a velocity of

only 0.73 m. per second at the altitude of 10 kilometers, and a velocity of 70 miles per hour would require, by the formula, an altitude of about 270 miles. This is not, therefore, an adequate cause of itself for the existence of such currents. But in addition to this there are the northeast and southeast trade-winds, extending up to a considerable altitude, and coming in obliquely from both sides toward the equator, or rather the doldrums a little north of the equator, and the deflecting force which produces the western component of velocity is counteracted by friction, except at and near the earth's surface, only after considerable westerly velocity is acquired; for friction can only be brought into play where there is a relative velocity between different strata, and so the velocity must increase with increase of altitude, and become large at high altitudes.

But, besides the preceding, there is another cause for westerly currents at high altitudes in the equatorial regions during the summer and winter seasons of the year. When both hemispheres have about the same temperature there is little or no interchange of air between them, and so, little or no current crossing the equatorial regions in either direction, either above or below. But at other times, during the summer of the northern hemisphere for instance, there is a current from the northern to the southern hemisphere above, and the reverse below, and it is readily seen that the current above, considered by itself and not in connection with the interchanging motions between the equatorial and polar regions at a mean annual temperature, in approaching the equator tends to acquire, by the deflecting force of the earth's rotation, a western component of motion, which continues to increase until the friction, which is small here, exactly counteracts this tendency, and mostly after the air passes the equator and the deflecting force changes to the left of the direction of motion, is this western component gradually decreased and at a short distance south of the equator changed to an eastern component of motion, though the effect of friction takes place in some measure before the air arrives at the equator, and while the deflecting force to the right, north of the equator, is vanishing. As the friction above is small,

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87. We have seen, § 77, that, according to theory, the atmosphere at the earth's surface has an easterly tendency in the higher latitudes, and the contrary in the lower ones. It must be evident to almost every one that in the middle and higher latitudes winds with a principal west component are both the strongest and most frequent, and that in the latitudes of the trade-winds the west component, especially upon the great oceans, is well marked, and without much variation. The few observations which we have from the interior of North Africa indicate that the prevailing winds are easterly and northeasterly. This is also shown to be the case on the northwest coast of Africa in the trade-wind latitudes by the sand and dust which is blown out into the Atlantic Ocean by the prevailing winds there. According to Mr. Laughton,[18] "The sand which these winds raise in the desert is carried by them far out to sea. During the summer it fills the air as far as the Canaries and to the height of the Peak of Teneriffe with an impalpable dust, which has the effect of a thick haze; and at all seasons of the year ships, even at a considerable distance from the shore, find it covering their sails and decks."

Professor Piazzi Smyth, also, observed on the Peak of Teneriffe "that the dust haze, which almost constantly filled the atmosphere and rendered the view very indistinct, was in the stratum of air moving from the eastward, that high overhead the air was clear, and that when the westerly wind descended to their level all traces of the dust haze in their immediate neighborhood vanished."[19] Of course these general tendencies are very much changed frequently by the numerous local and temporary disturbances, especially upon land, so that the directions, temporarily, are often nearly or quite reversed from that of the general tendency.

The average velocity of resultant motion of the air at the earth's surface for the whole year at a great many stations in all parts of the United States, according to the estimates of Coffin, is about two miles per hour with a resultant direction a little north of east. The east component of motion is, therefore, very nearly the same. These estimates, however, are extremely

vague and uncertain, being based mostly upon the relative frequency of the winds from the different principal points of the compass. Unfortunately the numerous observations of the Signal Service, as given in the Reports of the Chief Signal Officer, are of little avail in determining such resultants, since only the relative frequencies of direction from the different points are given, and not the corresponding velocities. The average hourly velocities, however, of motion in all directions are obtainable from the whole number of miles passed over in all directions by the air in the course of a month. From all the data of this kind available in the United States, and from similar data in Europe and Asia, Loomis[17] obtained the following average velocities in miles per hour without regard to direction:

Jan.	Feb.	Mar.	April	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Year.
United States (45 stations).												
9.69	10.13	10.97	10.13	8.88	7.95	7.28	6.91	7.97	8.90	9.83	9.75	9.5
Northern Europe (between parallels of 50° and 60°).												
11.23	11.54	11.45	10.15	9.84	8.99	9.01	9.19	9.06	10.60	11.52	11.21	10.3
Southern Asia, south of latitude 20° N. (34 stations).												
4.87	4.79	4.91	5.34	6.67	8.46	8.37	7.42	6.44	4.63	4.37	4.71	6.5
Near the West India Islands (7 stations).												
7.12	7.66	7.55	7.96	7.47	7.46	7.17	5.79	5.98	6.70	6.76	7.04	7.0

From these results alone we know nothing with regard to the resultants; but knowing that the general tendency of the air in northern latitudes is easterly, it is reasonable to suppose that the velocity of the east component of motion may be as much as two or three miles per hour.

At the West India Islands the wind blows pretty steadily in a direction nearly from the east, except when disturbed by cyclones, so that we may regard the average velocity of the west component as being at least five miles per hour. In southern Asia, on account of the great monsoon disturbances, little can be inferred from the results with regard to the velocity of this west component.

88. For theoretical reasons given in § 81, the velocity of the east component of motion at the earth's surface in the higher

latitudes of the southern hemisphere should be much greater than in the northern hemisphere. According to the universal experience of navigators, the westerly winds of these latitudes are nearly incessant, and very strong. With regard to the westerly winds of the higher latitudes of both hemispheres, and especially of the southern hemisphere, Mr. Laughton says:[18]

"In both hemispheres, to the north and south of the parallel of 35° or 40° a strong westerly wind blows with great constancy all around the world. In the southern hemisphere more particularly it blows with a persistency little less than that of the trade-winds, but with a strength which, although fitful, is very much greater. From a fresh strong breeze it rises frequently into a violent gale, and as such blows for days together; the mean direction being very nearly west, from which it seldom varies more than a couple of points on either side. South of the Atlantic, south of the Indian Ocean, south of Australia, in the higher latitudes of the southern Pacific, and to the southward of Cape Horn, we find it still the same—a westerly gale, whose strength and constancy combined have enabled Australian clippers to make passages which seem to border on the fabulous."

With regard to the winds of the southern hemisphere it is said:[19]

"The trade-wind system of the southern hemisphere is narrower than in the northern, extending from the equator to only about 30° and to about 33° in winter. The system of westerly winds comprises a zone bounded on the north by that of the equatorial winds just described, and extends southerly in the winter season to latitudes varying from about 61° in the southern Indian Ocean to about 65° or 66° in the South Pacific. Its southern limit in the summer season (their winter), which I have ventured to draw in latitude 50°, is as yet a matter of conjecture, no observations having been made from which to determine it.

"Of all the reliable resultants delineated in the well-defined portions of the system, amounting in the whole to over 250, every one in winter, and all but seven in summer, are westerly, and most of them are within a few degrees of the general mean, which is about N. 75° to 80° W."

89. With regard to the *annual inequality* in the easterly and westerly motions of the atmosphere, as deduced from theory, § 78, we have but few observations having any bearing on the subject, but these seem to indicate with some degree of

certainty that there really is such an inequality. It is observed upon the summits of Pike's Peak and Mt. Washington that easterly winds prevail more in the summer than in winter. Supposing the magnitude of the abnormal disturbances which entirely counteract and reverse the easterly motions so as to give rise to easterly winds to be the same at both seasons, it then follows that the easterly motions are less and so more frequently reversed in summer than in winter by these abnormal disturbances.

Away up in the region of the cirrus clouds we have seen, § 83, that the easterly velocities are so great that they are scarcely ever reversed there by abnormal disturbances, or even at lower altitudes in winter, for these easterly velocities at this season are unusually great. That the easterly velocities of the general motions of the upper strata are greater in winter than in summer may be inferred from observations by the Rev. Clement Ley upon the motions of the cirrus clouds. He says: "I have found that in winter very local depressions, even when deep, scarcely affect the directions of the cirrus currents in their vicinity, the latter continuing to be governed by the more general distribution of atmospheric pressures. Curiously enough this is not the case in summer." This would seem to indicate that the easterly velocities of the general progressive motions of the upper strata are so much greater in winter than in summer, that though they may be reversed by the abnormal gyratory disturbances in the latter season, they cannot be in the former.

On the earth's surface the velocities of resultant motions are so small, and these may be affected so much by monsoon influences, which have an annual period, that it is difficult to ascertain from observation whether there is an annual inequality in the general motions of the atmosphere at the earth's surface. In the estimates of the resultant motions at the earth's surface for numerous stations in the United States, Coffin found the velocities to be about twice as great in winter as in summer, and in a direction, in general, a little to the north of east. But this, being for only a comparatively small part of

the belt of easterly motions around the globe in the northern hemisphere, may be due to local circumstances of a monsoon character, such as the greater surface flow of air into the Atlantic Ocean in the winter than in the summer, since in the former season its temperature is greater, and in the latter less, than that of the continent.

It is seen from Professor Loomis's results, § 87, that in both the United States and in northern Europe there is an annual inequality in the velocities taken without regard to direction, and that, therefore, there is probably one in the velocities of the resultants, but this is not entirely conclusive, since the abnormal disturbances of a gyratory character may be greater in winter than in summer.

EFFECT OF THE GENERAL MOTIONS UPON ATMOSPHERIC PRESSURE.

90. Having investigated the effect of the temperature gradient in producing an interchanging motion between the equatorial and polar regions, and then the effect of these, in connection with the deflecting force of the earth's rotation, in causing east and west components of motion, the next step in order is to examine the effects of these latter motions, in connection with this same deflecting force, in causing gradients of atmospheric pressure, and so differences of pressure, between different parallels of latitude. While the atmosphere, if it had in all latitudes the same temperature, has no relative east or west motion, but only the absolute motion of rotation in connection with that of the earth's surface, the centrifugal force of rotation is just sufficient to cause the same ellipticity in the isobaric surfaces which the earth's surface has, and so to keep these surfaces parallel with the earth's surface. But if the atmosphere has an absolute east component of motion at all altitudes greater or less than that of the earth's surface, or, in other words, has an east or west component of motion relative to the earth's surface, then the atmospheric pressure at the earth's surface is no longer the same on all parallels of lati-

tude, and the isobaric surfaces are no longer parallel with the earth's surface, but pressure gradients with reference to this surface are produced in which the pressure either increases or decreases in a direction from the pole toward the equator, according as this component of motion is east or west. If the atmosphere at all latitudes had a relative east or west motion with a velocity on each parallel proportional to that of the absolute east velocity of the earth's surface, then the ellipticity of the isobaric surfaces would be increased or decreased just as much as that of the earth's surface would be if the angular velocity of the earth's rotation were changed by the same amount.

We have seen, § 71, that the effect of the upward expansion of the atmosphere due to the temperature gradient between the equator and the poles is to increase the ellipticities of the isobaric surfaces, and that this increase is in proportion to the altitudes of the strata, and therefore the east components of velocity, represented by v, which would give rise to a deflecting force sufficient to counteract the force arising from the gradient of this increased ellipticity, § 76, increases as the altitude. But since these, by means of the deflecting forces, are merely sufficient to keep the atmosphere, as it expands upward, from flowing away toward the poles, they have no effect upon the atmospheric pressure at the earth's surface. But if, now, the atmosphere at the earth's surface has an east or west component of motion of given velocity, that of all the strata above is changed by this same amount, and so diminished when the relative velocity at the surface is toward the west. The change of pressure, therefore, or pressure gradient, at the earth's surface depends only upon the east and west components of velocity at the earth's surface, and entirely vanishes where they vanish, whatever these velocities may be at altitudes above the earth's surface. At these altitudes, however, there are pressure gradients when there are no east or west components of velocity and pressure gradients at the surface; for since the deflecting forces arising from the east or west components of motion at different altitudes are just sufficient to

counteract those of the temperature gradients, these remain and, as we have seen, § 71, increase with increase of altitude.

What precedes is based upon the hypothesis that there is no interchanging motion of the atmosphere between the equatorial and polar regions—or at least, if there is such a motion, no sensible force or pressure gradient is required to overcome the frictional resistances to this motion and to maintain it. But we have seen that the east or west components of motion where there is friction cannot be maintained without the other; and so wherever there is polar motion, as in the upper strata, the east components of velocities must be a little less; and where there is equatorial motion, as in the lower strata mostly, they must be greater, or if the components of motion are toward the west, less, than those which would give rise to a deflecting force which would exactly counteract that of the temperature gradient.

91. The general expression of the barometric gradient G, corresponding to any given velocity s in any direction, so far as it depends upon the deflecting forces corresponding to this velocity, is given in § 57. In the special case now being considered, s becomes v, the east component of relative velocity of the atmosphere, and we therefore have here

$$G = 0.1571 v \sin l \frac{P}{P_0} \cdot \frac{T_0}{T}.$$

By substituting for v its general expression, § 77, in a function of the temperature gradient $\Delta\tau$ and of the east component of velocity at the earth's surface v_0, we get

$$G = 0.1571 (v_0 \sin l + 246.4 h \Delta\tau) \frac{P}{P_0} \cdot \frac{T_0}{T}.$$

From this expression of the barometric gradient we arrive at the same conclusion as in the preceding paragraph, namely, that the pressure gradient at the earth's surface, so far as it depends merely upon the east or west components of velocity at the earth's surface, vanishes where they vanish. For at the earth's surface we have h equal to 0, and as P there is generally

very nearly the same as P_0, we always have in this special case, very nearly,

$$G_0 = 0.1571 v_0 \sin l \frac{T_0}{T}.$$

It will be remembered that G is the change of atmospheric pressure in millimeters of mercury in the distance of one degree of a great circle of the earth, in a direction at right angles to the direction of motion, in case the rate of change should continue the same through so great a range. From this expression we may infer that in the northern hemisphere and in the higher latitudes, where v_0 is positive, the atmospheric pressure increases from the pole toward the equator to about the parallel of 35° or 30°, where v_0 vanishes and changes sign; but between this parallel and the equator, where v_0 is negative or west, the pressure decreases. In the southern hemisphere the gradient is reversed on account of the change of sign of sin l. Hence immediately south of the equator, where v_0 is negative, we have increasing pressure to about the parallel of 30°, where v_0 vanishes and changes sign, becoming positive again; after which the pressure decreases again to the pole, and very rapidly in the middle latitudes where the positive values of v_0 are very large (§ 88). At the equator, where sin l vanishes, the gradient also vanishes, and there is here consequently a minimum pressure. The pressure, therefore, in both hemispheres increases in the higher and middle latitudes in going from the poles toward the equator to about the parallel of 30° or 35° where there is a maximum pressure, and then decreases to the equator where there is a minimum. The gradients, however, in approaching the equator from either side, are small on account of the small value of sin l there, and consequently the equatorial depression is small. There is then, or would be if they were not interfered with by abnormal disturbances, a zone of high pressure in each hemisphere around the globe with its maximum near the parallel of 30°, an equatorial zone of low pressure with its minimum at the equator, and an area of low pressure around each pole with its minimum at the pole.

It seems difficult for some to conceive how the equatorial depression can be due to the westerly motion of the atmosphere, since this motion prevails at the earth's surface and in the lower strata of the atmosphere only, while above the motion is easterly. But it has been shown that in the case of no east or west component of motion at the earth's surface the east components are necessary to keep the air above from flowing away from the equatorial region, and thus from diminishing the pressure there. If, then, the lower stratum next the earth's surface has a west component of motion, we have seen that the velocities of all the strata above to the top of the atmosphere have their velocities changed by the same amount, and the effect upon the pressure at the earth's surface is precisely the same as if there were no temperature gradient and no east components of velocity above, and all the strata from bottom to top should receive a west component of motion.

92. According to the preceding approximate general expression of G, it is seen that the pressure gradient at the earth's surface, where $h=0$, is positive where $v_0 \sin l$ is, and vanishes and changes sign at the latitude where v_0 does. The greatest pressure, therefore, at the earth's surface is on this parallel, or nearly. But at any altitude above the earth's surface the expression of G vanishes at some latitude nearer the equator, since after v_0 becomes negative, the term in the parenthesis depending upon h being always positive in the northern hemisphere, and in either of a contrary sign to that of $v_0 \sin l$ in these latitudes, the expression only vanishes where the negative value of $v_0 \sin l$ is equal to the part in the parenthesis depending upon h, and hence the greater h is, the nearer the equator is the latitude where G vanishes and the pressure is a maximum. In order to make G vanish, the following conditions must be satisfied:

$$v_0 \sin l + 246.4 h \Delta\tau = 0.$$

The altitude on any given latitude where this condition is satisfied of course depends upon the value of v_0, but for some distance toward the equator from the parallel where v_0 vanishes, the term $v_0 \sin l$ increases although $\sin l$ diminishes; and hence

the greater the altitude the nearer the equator is the latitude where G vanishes and the maximum pressure occurs in going horizontally toward the equator. The values of v_0, however, corresponding to the different latitudes or values of sin l, may be so small that at and above a given altitude the second term above may, for all latitudes, be greater than the negative value of the first one, and then the expression is positive and does not vanish at all, except at the equator, where both sin l and $\Delta\tau$ vanish; and therefore, above a given altitude and at all higher altitudes, the maximum pressure, in going horizontally toward the equator, is only reached at the equator, and at these altitudes there is a continuous gradient of pressure increasing from the pole to the equator. With no value of v_0 on any parallel, that is, with no east or west components of motion at the earth's surface, we have seen there would be no pressure gradient at the earth's surface, so far as it depends upon v and the deflecting forces, but at any altitude above the surface there would be a gradient of increasing pressure from the pole to the equator, and this at great altitudes is so large, that the west component of motion of the atmosphere at the earth's surface, and the consequent change of velocities by the same amount at all altitudes, are not sufficient, by means of the deflecting forces, to reverse this gradient except in the lower strata of the atmosphere up to a certain altitude. Above this altitude, therefore, there is no barometric minimum at the equator, but a maximum, arising from the gradual nearer approach of the maxima on each side towards the equator as the altitude increases.

It must be always borne in mind that the preceding relations are not strictly correct, but only approximately so, since the value of v, § 77, which has been used in obtaining the preceding expression of G is approximate only, being the true expression in the case of no friction, in which case there is no necessity for an interchanging motion between the equator and the poles, but is in all cases a limit beyond which the value of v cannot go and of which it must fall a little short, and the greater the amount of friction to be overcome the more so.

93. From what precedes, there is, at the earth's surface, an area of low pressure around each pole with its minimum at the pole, a zone of high pressure in each hemisphere with its maximum about the parallel of 30°, and an equatorial zone of slightly diminished pressure with the minimum at the equator. And this same arrangement exists in the lower strata of the atmosphere up to a considerable altitude,—except that the polar depressions are greater and the parallel of maximum pressure comes nearer the equator as the altitude is increased, until at a certain altitude the two maxima combine to form a maximum pressure at the equator. Instead, then, of the isobaric surfaces being elliptical, and having an increasing ellipticity in proportion to the altitude, as represented in Fig. 3, they have, in consequence of the east and west components of motion of the atmosphere at the earth's surface, steeper gradients in the higher and middle latitudes, a bulging up nearer the equator, and in the lower strata a minimum altitude at the equator, as represented in Fig. 4, in which the intersections of the line *eo* with the isobaric surfaces indicate the latitudes, at the several altitudes, where the pressures are the greatest and where the east components of velocity vanish, change sign, and become west components of velocity. The actual form and position of the line *eo*, since they depend upon the west components of motion at the earth's surface, and the temperature gradient, are quite uncertain toward the equator.

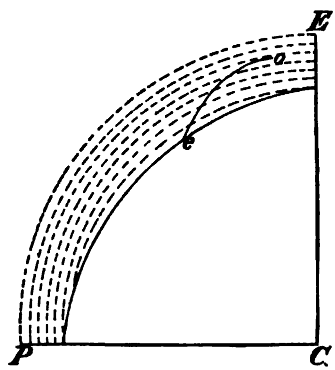


Fig. 4.

Before the exploring expeditions led by Captain Wilkes and Sir James Ross it was pretty generally thought that the barometric pressure at sea-level in all parts of the earth is nearly or quite the same, and about 30 inches. From the barometric observations of these expeditions it was first clearly

shown that this is by no means the case. Says Captain Wilkes:[35]

"The most remarkable phenomenon which our observations have shown is the irregular outline of the atmosphere surrounding the earth as indicated by the pressure upon the measured column at different parts of the surface. Our barometrical observations show a depression within the tropics, a bulging in the temperate zone, again undergoing a depression on advancing towards the arctic and antartic circles."

Says Sir James Ross:[36]

"Our barometrical experiments appear to prove that the atmospheric pressure is considerably less at the equator than near the tropies; and to the south of the tropic of Capricorn, where it is greatest, a gradual diminution occurs as the latitude is increased, as will be shown from the following table, derived from hourly observations of the height of the column of mercury between the 20th of November, 1839, and the 31st of July, 1843."

EXTRACT FROM ROSS'S TABLE.

Latitude.	Pressure, Inches.	Latitude.	Pressure, Inches.	Latitude.	Pressure, Inches.
Equator.	29.974	42° 53′	29.950	55° 52′	29.360
13° 0′S.	30.016	45 0	29.664	60 0	29.114
22 17	30.085	49 8	29.467	66 0	29.078
34 48	30.023	51 33	29.497	74 0	28.928
		54 26	29.347		

So strange and unaccountable did these phenomena appear before the discovery of the deflecting force of the earth's rotation, that Espy said: "Unless I have positive testimony to the fact, I shall not believe that the barometer stands lower at the level of the sea in the southern hemisphere than in the northern, *only in regions where great snows or rains prevail, and only while they do prevail.*" This was upon the theory that great barometric depressions are caused by a great quantity of aqueous vapor in the atmosphere.

94. Since the whole vertical circulation and the east and west components of motion, and the deflecting forces arising from them, from which result the polar and equatorial depres-

sions and the zones of high pressure near the tropics, depend upon the temperature gradients between the equator and the poles, there must be greater polar depressions of the isobaric surfaces and a greater bulging up in the middle and lower latitudes in winter than in summer, since the temperature gradients are much greater in the former season than in the latter, especially in the northern hemisphere, where they are more than twice as great in January as in July. There must, then, be an annual inequality of atmospheric pressure from this cause, such that the pressure is greater in the winter than in the summer of each hemisphere in the middle and lower latitudes, and the reverse in the polar regions.

But for another reason, also, there is an annual oscillation of pressure. During the winter of each hemisphere the atmosphere becomes much colder than it is in the other hemisphere, and consequently its volume considerably less, so that the isobaric surfaces lie lower in the colder hemisphere than in the other. There is, consequently, a pressure gradient above by which the air of the higher strata flows from the warmer hemisphere to the colder one and increases the mass and pressure of the atmosphere there a little, and diminishes them by the same amount in the warmer hemisphere, until there is a reverse gradient formed in the lower strata which gives rise to a counter flow of air below from the colder to the warmer hemisphere. There is, in this way, an interchanging motion originated and maintained between the two hemispheres, just as there is between the warm equatorial and cold polar region of each hemisphere, as has been described in § 72. But the interchange between the two hemispheres is kept up with greater facility, since there are no east components of motion at and near the equator, where the temperature and pressure gradients in this case and the velocity of flow are the greatest, to give rise to counter deflecting forces, as there are in the middle latitudes in the other case, but the whole temperature and pressure gradients are brought to bear in maintaining the flow of air above from the warmer to the colder hemisphere, and the contrary in the lower strata.

As there is an annual inversion of the temperature conditions which give rise to and maintain these interchanging motions between the two hemispheres, so there is also one in the pressure gradients at the earth's surface and in the lower strata, arising from the flow of air above from the warmer to the colder hemisphere, and in the slight increase of pressure there and corresponding diminution in the other hemisphere, to cause this gradient. There is therefore a little more air, and consequently a little greater pressure at the earth's surface, in each hemisphere in winter than in summer, and this causes a slight annual inequality of pressure, the pressure being the greatest in each hemisphere in the midwinter of that hemisphere. The greatest effect of this latter cause of the annual inequality is at the poles, whereas in the other case the pressure is increased in winter in the middle latitudes, but diminished in the polar regions. It is the resultant of these two effects which is observed.

95. An annual inequa'ity of pressure is observed on nearly every part of the globe, but on account of various local disturbances of pressure, this, as the temperature, is not the same at all places on the same parallel. But by obtaining the normal pressures of latitude, as those of temperature, by taking the average of observed values all around the globe on each parallel, we eliminate the effects of these local disturbances. In this way the results given in the following table have been obtained.[8] These results, however, must be regarded as being only approximate, since the material on hand then, more than ten years ago—was insufficient for accurate results—since there was a great deficiency of observations over nearly all parts of the great oceans. It is much to be regretted that the work has not been repeated with the now much greater material on hand for such a purpose.

In the following table are given for the earth's surface the normal barometic pressure of latitude, P; the barometric gradient, G; and the annual variations, ΔP and ΔG, for each fifth degree of latitude, so far as there were any available observations.

Latitude.	Annual Mean.		Annual Inequality.	
	P	G	ΔP	ΔG
	mm.	mm.	mm.	mm.
+80°	760.5	....	−0.06	....
75	760.0	−0.19	+0.19	+0.04
70	758.6	−0.14	0.36	0.05
65	758.2	+0.01	0.63	0.06
60	758.7	0.15	0.97	0.06
55	759.7	0.20	1.26	0.05
50	760.7	0.18	1.41	0.03
45	761.5	0.15	1.53	0.02
40	762.0	+0.07	1.61	+0.01
35	762.4	−0.03	1.66	0.00
30	761.7	0.18	1.66	−0.01
25	760.4	0.25	1.61	0.03
20	759.2	0.21	1.41	0.06
15	758.3	0.13	1.05	0.09
10	757.9	−0.03	+0.50	0.11
+5	758.0	+0.01	−0.05	0.12
0	758.0	0.04	0.63	0.12
−5	758.3	0.11	1.18	0.11
10	759.1	0.20	1.70	0.08
15	760.2	0.26	2.00	0.06
20	761.7	0.29	2.22	−0.03
25	763.2	+0.18	2.36	0.00
30	763.5	−0.08	2.22	+0.03
35	762.4	0.30	1.85	0.06
40	760.5	0.51	1.41	0.07
45	757.3	0.73	1.00	0.09
50	753.2	0.91	−0.50	0.10
55	748.2	0.97	0.00	+0.10
60	743.4	0.83		
65	739.7	−0.56		
−70	738.0			

From the preceding table it is seen that the mean normal pressure of latitude is greatest in the northern hemisphere about the parallel of 35°, and in the southern hemisphere a little nearer the equator, about the parallel of 30°, and that there is an equatorial depression and a polar one in each hemisphere, in accordance with the theoretical deductions of § 93. The mean pressure gradient is greater in the southern than in the northern hemisphere, and is especially large in the vicinity of the parallel of 55° S., as it should be according to theory and the theoretical expression of G, § 91, since the east and west components of velocity, or values of v_{θ}, are greater in the

former than in the latter, and the east component is especially large at and near the parallel of 55° (§ 88). The pressure gradients, being reckoned on the meridian from the north toward the south pole, have contrary signs generally on corresponding parallels of the two hemispheres.

96. The coefficients of the annual inequalities ΔP and ΔG in the preceding table, added to the mean pressure and mean pressure gradient for the year, give the mean pressure and mean pressure gradient for January, and subtracted, those for July. It is seen, therefore, that the pressures, by observation, are greatest in each hemisphere during the winter of that hemisphere, except very near each pole, as required by theory, § 94. The annual inequality of pressure, by the theory, is the result of two separate effects, which in the middle and lower latitudes of each hemisphere are both in the same direction, tending to increase the pressure in the winter season of the hemisphere. But the effect which arises from the bulging up of the isobaric surfaces in the middle and lower latitudes, and a corresponding depression in the polar latitudes, has a contrary sign in the polar latitudes, and seems to be of such a magnitude as to more than counteract the other effect, and hence in these latitudes the annual inequality is reversed, and the barometric pressure is less during the winter than the summer season. There is, consequently, a parallel very near the pole in the northern hemisphere, but near the parallel of 55° in the southern, where there is no annual inequality of barometric pressure.

The annual inequality of pressure is greater in the southern than in the northern hemisphere, and the maximum nearer the equator, being in the southern hemisphere on the parallel of 25°, while in the northern it is about the parallel 32°.5. This arises from the greater amount of water surface and consequently less amount of frictional resistance to the east and west components of motion at the surface in the southern hemisphere than in the other. For since, on this account, there is a greater mean bulging up of the strata of equal pressure in the middle and lower latitudes, and depression of them in the

polar latitudes, corresponding to the mean temperature gradients of the year, in the southern than in the northern hemisphere, the annual inequality of pressure corresponding to the annual inequality of temperature must likewise be greater in the former than in the latter, for the effect of the annual reversion of the interchanging motions between the two hemispheres, upon the east components of motion of the atmosphere, and consequently upon the pressures, must be the greater in the hemisphere which has the more water surface.

EAST VELOCITIES DEDUCED FROM PRESSURE AND TEMPERATURE GRADIENTS.

97. To any pressure gradient G at the earth's surface, or any altitude above it, there is a corresponding velocity v of east or west component of motion, § 91, so that if one is known the other may be approximately computed. We can more easily obtain from observation the normal gradients of pressure than the east or west component of wind velocity, and therefore this component can be more readily obtained from the observed pressure gradients than directly from observation. For instance, taking the parallel of 50° N. for an example, we find in the preceding table the normal gradient of pressure at the earth's surface is 0.18 mm. With this value of G and the value of sin 50°, Table V, we readily get from the special expression of G_0 in § 91, $v_0 = 1.5$ m. as the east component of velocity at the earth's surface. Now from the temperature differences for 10° preceding and following the latitude of 50° for the mean of the year, § 68, we get approximately the gradient $\Delta\tau = (7.9 + 7.3)/20 = 0.76$. This is for the distance of one degree, in which case the numerical coefficient 0.00221 must be used in the expression of v in § 77. With these values of v_0 and $\Delta\tau$ this expression gives for the altitude $h = 1000$ meters. $v = 1.5 + 2.2 = 3.7$ m. for the value of the east component of motion, in meters per second, at the height of one kilometer. For any other altitude the last term would be increased in proportion. Thus, for an altitude of 5 kilometers we should have

$v = 1.5 \times 2.2 \times 5 = 12.5$ in meters per second. Multiplying by 3.6, we get $v = 45$ in kilometers per hour.

98. Somewhat in the same manner, the following table of values of v in kilometers per hour have been computed for the mean of the year, and for January and July, for each fifth degree of latitude, except near the equator and the poles, where the formula and the data are too uncertain. The temperature gradients, however, were obtained by another method, more accurate than that used above.[*]

Latitude.	v		
	Mean of the Year.	January.	July.
	km.	km.	km.
+75	−4.4+4.8 h	− 1.9+4.4 h	−5.7+5.1 h
70	−3.3 6.6 "	− 1.1 7.5 "	4.8 5.8 "
65	+0.2 7.7 "	+ 1.6 9.6 "	−1.2 5.9 "
60	3.9 8.3 "	5.5 10.9 "	+2.3 5.6 "
55	5.5 8.5 "	7.0 11.6 "	4.0 5.4 "
50	5.4 8.6 "	6.4 12.1 "	4.4 5.1 "
45	4.9 8.8 "	5.5 12 6 "	4.1 5.0 "
40	+ 2.6 9.0 "	+2.8 12.9 "	+2.4 4.8 "
35	− 1.2 9.3 "	−1.0 13.8 "	−1.4 4.6 "
30	8.6 9.5 "	9.1 14.7 "	8.1 4.3 "
25	14.4 9.4 "	16.0 15.2 "	12.2 3.6 "
20	15.1 9.0 "	20.2 15.6 "	11.7 2.4 "
+15	12.5 5.6 "	21.8 10.5 "	3.8 0.6 "
......	....	....	....
......	....	....	....
−15	25.0 8.2 "	18.7 4.9 "	31.0 11.7 "
20	20.9 7.8 "	18.8 5.8 "	23.0 10.0 "
25	− 10.3 7.6 "	−10.4 6.6 "	−16.1 8.7 "
30	+ 3.8 7.5 "	+ 2.3 7.3 "	+ 5.2 7.8 "
35	12.4 7.4 "	10.0 7.8 "	14.4 7.1 "
40	18.7 7.4 "	16.0 8.2 "	21.4 6.7 "
45	24.0 7.4 "	21.0 8.6 "	27.0 6.4 "
50	27.5 7.5 "	24.5 8.9 "	30.5 6.1 "
55	27.3 7.5 "	+24.6+9.1 "	30.0 5.9 "
60	+21.9+7.5 "		

By adding together the two parts of the value of v, the first being v_0, the value of v at the earth's surface, where $h = 0$, and depending upon and deduced from the pressure gradient there, and the second depending upon the altitude h, we get the east component of velocity v in kilometers per hour at any given altitude h in kilometers. But by reducing the first part to

miles by multiplying by 0.62 we get the result in miles per hour if the value of h is given in miles. For instance, on the parallel of 50° N. and for altitude $h = 5$ kilometers, we have for the mean of the year $v = 5.4 + 8.6 \times 5 = 48.4$ kilometers per hour. This differs a little from the result of the preceding computation with the temperature gradient deduced approximately from first differences merely where the intervals are ten degrees of latitude. In like manner, by reducing the first part to miles, we have $v = 3.3 + 8.6 \times 5 = 46.3$ miles per hour for the east component of velocity at the altitude $h = 5$ miles.

In the same way we get, on this same parallel of latitude, $v = 4.0 + 12.1 \times 5 = 64.5$ miles per hour for the approximate east component of velocity of the air at the altitude of 5 miles in January, and $v = 2.8 + 5.1 \times 5 = 28.3$ miles per hour for this velocity in July. Hence the east component of velocity at considerable altitudes is more than twice as great in January as in July on this latitude, and the same is true of the middle and higher altitudes generally in the northern hemisphere, as has already been deduced in a general way in § 78. At and near the earth's surface, however, where the value of v depends mostly upon v_0, and especially in the lower latitudes where the value of v_0 is considerable and negative, the ratio between January and July is a little different, since it depends mostly or entirely upon that of v_0 for the two extremes of the seasons, in which the uncertainty of friction corresponding to different velocities comes in, so that this part, as deduced from the barometric gradients of pressure, does not quite have the same relations between the different seasons as the temperature gradients.

In the lower latitudes, where there is a west component of motion at the earth's surface, we have v_0 negative, and so these west components of velocity are diminished above by the quantity, for the different altitudes, in the preceding table depending upon h, so that at certain altitudes they become reversed and become east components of velocity. Thus on the parallel of 20° N. we have, for the mean of the year, $v = -15.1 + 9.0h$. It is readily seen that in order to make v van-

ish we must have $h = 15.1/9.0 = 1.7$ km. nearly for the altitude where v vanishes and changes sign, and where, consequently, there is no east or west component of motion of the air. It is seen from the table that this altitude differs for different parallels of latitude and the different seasons of the year, being very small at the parallel of 30° and gradually increasing toward the equator, and being also, at and within the tropics, very much greater in July than in January.

In the southern hemisphere on the parallel of 50° we have for the mean of the year at the altitude of 5 kilometers, $v = 27.5 + 7.5 \times 5 = 65$ kilometers for the east component of velocity of the air per hour. This is considerably greater than that obtained above for the same latitude and altitude in the northern hemisphere, due to the greater east component of velocity at the earth's surface for reasons already explained. The differences between January and July, as given by this table, are very small in comparison with those of the northern hemisphere on account of the comparatively small differences, generally, in the temperatures, as may be seen from the table of § 68.

99. We have seen that the maximum atmospheric pressure, § 92, in going at any altitude horizontally toward the equator, occurs where $v = 0$; and hence the latitude of greatest pressure at any given altitude is readily ascertained approximately from a mere inspection of the preceding table, it being on the parallel which makes the expression of v vanish at the given altitude. Thus, in the northern hemisphere, for the mean of the year, at the altitude of one kilometer, the latitude on which this condition is satisfied is a little below 30°, while at the earth's surface the greatest pressure is about 36°. At greater altitudes this latitude is still nearer the equator, but the exact latitude becomes more uncertain on account of the uncertainty of the data, and the small changes of pressure with a given change of latitude at these altitudes. Above a certain altitude, it is readily seen, so far as the table extends, that the condition cannot be satisfied, and hence east velocities prevail above this altitude.

From the applications which have been made of the preceding table, both to the east and west component of velocity

of the air and to the pressures, it is seen that this table is of great importance in studying the general motions and pressures of the atmosphere on different parallels of latitude and at different altitudes, and also their annual changes with the seasons of the year, although the results are to be regarded as being only approximate, and, near the equator, not at all accurate.

SURFACE WINDS, CALMS, AND CALM BELTS.

100. So far we have considered mostly the general circulation of the great mass of the atmosphere generally, and determined the effect of this in connection with the deflecting forces of the earth's rotation, upon the atmospheric pressures. We come now to consider the reflex and secondary action of this modified pressure in modifying the general circulation arising directly from the temperature gradients between the equator and the poles and the deflecting forces. It has been shown, § 93, that in this modified pressure there is a zone of high pressure in each hemisphere, with its maximum pressure near the parallel of 30°, and extending all around the globe. The effect of this is to cause the air at the earth's surface to flow out from beneath, on the one hand toward the equator, and on the other toward the pole. In the former case the flow of air at the surface, combining with the general flow in the lower strata from the pole toward the equator, and acted upon by the deflecting force of the earth's rotation, adds greatly to the strength of the trade-winds. In the latter case, the flow being in a direction contrary to that of the general flow of the lower strata from the pole toward the equator, and being also a little stronger at and near the earth's surface, there is a residual motion toward the pole in the middle latitudes, which, also acted upon by the deflecting force of the earth's rotation, inclines toward the east, and gives rise to the gentle southwest winds in the northern and northwest winds in the southern hemisphere in these latitudes.

It must be borne in mind in this connection that we are not here considering the actual circulation and other conditions of

the atmosphere as disturbed from various abnormal causes, but the case of an earth with a homogeneous surface, and with uniform temperatures at all longitudes on the same latitudes. To the results here obtained must still be added those depending upon the irregular and abnormal disturbances in order to get the true results as observed.

The effect of these zones of high pressure, with their maxima in the lower part of the atmosphere nearer the equator as the altitude is increased, is undoubtedly felt to a considerable altitude, but mostly near the earth's surface. Above, where there is little friction, the equatorial and polar motions arising from pressure gradients, however produced, we have seen give rise to east or west components of motion which, by means of the deflecting forces of the earth's rotation, almost entirely counteract the effect of the pressure gradients, and leave the velocities of the polar and equatorial motions very small in comparison with what they otherwise would be, except very near the equator, where the deflecting forces are very small. But near the earth's surface, where the friction is great, it requires greater polar and equatorial components of motion to overcome the friction and to keep up east and west components of velocity which, by means of the deflecting forces arising from them, would materially interfere with the free outflow below. Besides, since this outflow of air below from beneath the high pressure is confined mostly to a comparatively thin stratum at the earth's surface, there is great resistance to this flow both from friction between it and the earth's surface and from the strata above, which have much less motion, and this in a somewhat different direction, since the motion above is more nearly in an east or west direction. It is therefore necessary to have comparatively small east or west components of velocity, so as to leave nearly the full force of the pressure gradient to overcome the resistance to the polar and equatorial motions near the earth's surface. For two reasons, therefore, it is necessary here, and wherever there is much friction, that the polar and equatorial components of motion shall be large in comparison with the east and west ones. While therefore the motions of

the air are nearly in a direction either east or west on each side of the zone of high pressure at a small elevation above the earth's surface, at and near the surface they deviate considerably from these directions, and assume more nearly polar and equatorial directions. Looking at the matter in a more general way, all the conditions of the general atmospheric circulation, in the case of no friction, are satisfied with east and west motions only; for the initial motions may be such that the deflecting forces of these motions exactly counteract the forces arising from the temperature and pressure gradients. But the greater the friction and the less the deflecting forces of the earth's rotation, the greater must be the equatorial and polar motions in comparison with the others, and the greater the deviation of the resultant from an east or west direction. As the deflecting force of the earth's rotation becomes small on the trade-wind parallels, the trade-winds have a smaller west component of motion than they otherwise would have in comparison with the other component.

101. It is probable that the polar motion of the lower part of the atmosphere in the middle latitudes does not extend, so as to be sensible, beyond about the parallel of 60° in either hemisphere. In the northern hemisphere there are too many abnormal disturbances for the determination of this limit. In the southern hemisphere, on account of the smooth-water surface in the middle and polar latitudes, there are fewer disturbances of this sort; but here there is a great deficiency of observations: the few, however, which have been made indicate that in the polar region south of the parallel of 60° or 65° there is an equatorial component of motion. With regard to the winds of this region Coffin says:[18]

"Of the third system, comprising the southern polar winds, our knowledge is confined to the months of winter and early spring (of the northern hemisphere). Of the five resultants near the margin of the zone, 60° to 65°, all are from nearly a south point. Of the whole fifteen within the zone only five tend toward, and two of them but slightly so, while ten recede from it, generally at a large angle."

It seems, therefore, that in the polar regions of this

hemisphere the winds, upon the whole, have a component of motion from the pole.

102. It has been shown, § 78, that the east component of velocity of the air at the earth's surface in the middle and polar latitudes depends upon the rapidity of the vertical circulation and the relative motions of the strata with reference to one another at different altitudes. But these relative motions, or differences of motion of different strata, depend upon the deflecting forces of the earth's rotation; and where these forces are very weak, they do not nearly come up to the limit given by the expression of v in § 77, which gives the limit beyond which the value of $v-v_0$ cannot go, and consequently to the relative velocities between the strata having an east component of velocity v_0 at the earth's surface, and v at any altitude h.

The forces deflecting towards the east or west depend upon the horizontal polar or equatorial velocity, as is seen from the expression of F_u, § 49. But very near the poles the value of u nearly vanishes, since the interchanging motion of any particle of air between the equatorial and polar regions is oscillatory, having its greatest velocity in the middle latitudes, which gradually decreases both toward the equator and the poles, and nearly vanishes just before it arrives at its greatest limit of range. Near the pole, therefore, there is not much difference between the east components of velocity of the upper and lower strata, and consequently little force to overcome the friction at the earth's surface and keep up an east component of velocity there. Since, therefore, this component of velocity in the polar regions is small, and also the component of motion from the poles, these must be regions in which there is very nearly a calm, unless there is some abnormal disturbance.

At and near the equator the deflecting forces which cause a difference in the east and west components of velocity at different altitudes vanish, or nearly so, for two reasons: the one, because as near the pole, and for the same reason, there is little motion either toward or from the equator; and the other, because the deflecting forces here, with any given velocity u of motion toward or from the equator, decrease as sin l de-

creases, and entirely vanishes at the equator. The relative velocities between the strata, therefore, vanish at, and are very small near, the equator; and consequently the same is the case with the force which overcomes the friction at the earth's surface and maintains an east or west component of motion there. And as there is no motion north or south between the systems of winds of the two hemispheres where the trade-winds meet, which is near the equator, there is here a belt of calms extending around the globe, where not interfered with by abnormal disturbances or forces not considered in the general circulation of the atmosphere.

103. For reasons given in § 80 there can be no east or west component of motion at the earth's surface on the parallel of maximum atmospheric pressure on the earth's surface, at or near the parallel of 30°. And as the air is pressed out at the earth's surface, toward the pole on the one side and the equator on the other, from beneath the zone of high pressure, there is no sensible motion north or south in the middle part of this zone. There is then here, in each hemisphere, a belt of calms, extending around the globe, and having its middle on this parallel of latitude, except so far as it is interfered with by abnormal disturbances not here considered.

Since the deflecting forces arising from the much greater east components of velocity in the southern hemisphere, both at the earth's surface and at all altitudes, are much stronger than the similar and opposing forces of the northern hemisphere, the southern system of circulation encroaches somewhat upon the area of the other, causing all the calm-belts to be a little further north than they otherwise would be, so that the equatorial calm-belt is a little north of the equator, while the tropical calm-belt of the northern hemisphere is a little further from the equator than that of the southern hemisphere. For the same reason there is, on the average of the year, a little more air in the northern than in the southern hemisphere, as may be seen from an inspection of the barometric pressures in the second column of the table of § 95, although the maximum about the parallel of 30° in the southern hemisphere is a

little greater than that of the northern, on account of a little greater bulging up of the isobaric surfaces.

SUMMARY AND GRAPHIC REPRESENTATION OF THE MOTIONS AND PRESSURES.

104. In the preceding part of this chapter it has been shown that if all parts of the atmosphere had the same temperature there would be a complete calm over all parts of the earth's surface. But that in consequence of the difference of temperature between the equatorial and polar regions of the globe, and the consequent temperature gradient, there arise pressure gradients and forces which give rise to and maintain a vertical circulation of the atmosphere with a motion of the air of the upper strata of the atmosphere from the equator toward the poles, and a counter current in the lower part from the poles toward the equator, as represented by the arrows in the following figure, and that this of course requires a gradual settling down of the air from the higher to the lower strata in the middle and higher latitudes, and the reverse in the lower latitudes. It has also been shown that in case the earth had no rotation on its axis, this would be exclusively a vertical circulation in the planes of the meridians without any east or west components of motion in any part; but that, in consequence of the deflecting forces arising from the earth's rotation, the atmosphere at the earth's surface has also an east component of motion in the middle and higher latitudes, and the reverse in the lower latitudes, and that the velocities of the east components increase with increase of elevation, so that at great altitudes they become very much greater than those at the earth's surface; while those of the west components decrease with increase of altitude up to a certain altitude, where they vanish and change signs and become east velocities, now increasing with increase of altitude to the top of the atmosphere.

It has been further shown that the deflecting forces arising from the east components of motion of each hemisphere from the earth's surface to the top of the atmosphere, in the middle

and higher latitudes and of the upper part of the atmosphere in the lower latitudes, drives the atmosphere from the polar regions toward the equator, while those arising from the west components of motion in the lower part of the atmosphere in the lower latitudes, having a contrary effect, but small in comparison with the other on account of the weakness of these forces near the equator, tend to drive the air a little from the equator toward the poles. There is, therefore, a depression of the isobaric surfaces at all altitudes in the polar regions, especially in the southern hemisphere, a much smaller depression in the equatorial regions, and a bulging up of the isobaric surfaces in the vicinity of the parallel of 30° in the lower part of the atmosphere, the maximum being nearer the equator as the altitude increases, as represented in Fig. 4, but at high altitudes there is a minimum of barometric pressure at the poles and a maximum at the equator.

Fig. 5.

In the accompanying figure the solid arrows in the interior part represent the resultant motions of the winds (longer ar-

rows indicating greater velocities), in case of an earth with a homogeneous surface over both hemispheres, in which the motions would be symmetrical in both and the same at all longitudes, and the equatorial and tropical calm-belts would be situated at equal distances from each pole. The dotted arrows indicate the strong, almost eastern motion of the air at all latitudes at some high altitude, as that of the cirrus clouds.

The outline of the outer part of the figure represents an isobaric surface high up where the bulging up near the parallel of 30° disappears and the maximum pressure at the same altitude is transferred to the equator. For lower altitudes the isobaric surfaces have a bulging up at the parallel of 30°, and a slight depression at and near the equator. The arrows in this part represent the polar and equatorial components of motion, the former above and the latter below, except near the earth's surface on the polar sides of the tropical calm-belts, where there is a polar component of motion arising from the air's being pressed out from under the belt of high pressure. This, perhaps, does not extend beyond the polar circles, beyond which there can be little motion in any direction, except from abnormal disturbances.

For reasons given in § 103, the actual mean positions of the equatorial and tropical calm-belts are not precisely as here represented, but are all a little displaced toward the north pole, and the polar depression of the isobaric surfaces is greater in the southern than in the northern hemisphere.

ANNUAL OSCILLATION OF THE CALM-BELTS.

105. For the same reason that there is a semi-annual inversion of the interchanging motions between the two hemispheres, in the one direction in the spring and the contrary in the fall, from which results an annual inequality of the atmospheric pressure at the earth's surface, § 94, there is also an annual oscillation of the calm-belts. During the summer of the northern hemisphere, while there is a flow of air below from the

southern to the northern hemisphere and the contrary above, the vertical circulation and interchanging motion of the southern hemisphere, due to the difference of the mean temperatures of the equatorial and polar regions, is strengthened while that of the northern hemisphere is weakened. The consequence is that the southern stronger system at this season encroaches somewhat upon the territory of the other, causing the middle of the equatorial calm-belt, which is the dividing line between the two systems, to be a little north of its mean position; and as the tropical calm-belts are intermediate between the equatorial calm-belt and the poles, and their positions must depend somewhat upon the extension and the limits of the systems respectively to which they belong, the positions of these also must be thrown further north, when that of the equatorial calm-belt is. Of course just the reverse takes place during the winter of the northern hemisphere. There is, therefore, an annual oscillation of all the calm-belts, such that they have their most northerly position in midsummer and the reverse in midwinter of the northern hemisphere.

Or looking at the matter in a somewhat different way, we have seen that the southern system of circulation, for the mean of the year, especially the east components of motions, due to the same temperature gradients, or nearly, between the equator and the poles as in the northern hemisphere, is stronger than that of the northern one on account of there being less frictional resistance at the earth's surface to the former than to the latter, and that in consequence of this the mean position of the equatorial calm-belt is a little north of the equator and both the tropical calm-belts are a little farther north than they otherwise would be, and so at unequal distances from the equator. Now the effect is the same whether we diminish the resistance or increase the forces. In the summer season, therefore, of the northern hemisphere, when the forces are increased in the southern and diminished in the northern hemisphere, we have exactly the same effect as where the resistances are less in the southern and greater in the northern hemisphere, namely,

that all the calm-belts are thrown a little to the north of their mean positions, or positions which they otherwise would have; and the reverse in the winter season of the northern hemisphere.

106. Since the equatorial calm-belt depends upon two circumstances, the meeting below of the air currents belonging to the two hemispherical systems of circulation, and the vanishing of the forces which cause a westerly motion of the air, its northern limit is better defined than the equatorial or southern limit, and its range of oscillation is greater. In the summer season of the northern hemisphere, when the lower currents of the two system meet at a greater distance north of the equator, the forces on the north side of the belt which gives rise to the west component of motion are very much stronger than those on the equatorial side very near the equator, where these forces almost entirely vanish. Over the whole space therefore between the calm-belt and the equator these forces sensibly vanish, so that, although the southeast trade-winds blow beyond the equator, yet they have here, and for a few degrees north of the equator, no west component of motion, and become southerly winds, and of little strength on account of their being so near the place of meeting where all motion ceases. The winds, then, between the equator and the calm-belt are weak, especially at the season when this belt is near the equator, so that the southern limit of the belt is then uncertain, and the southeast trade-winds appear to vanish soon after passing the equator, and even when the belt has its most northerly position these winds are too weak to be observed far beyond the equator. There is, therefore, an apparent widening of the calm-belt during the summer season of the northern hemisphere.

107. The preceding theoretical deductions with regard to the calm-belts are fully confirmed by observations of the trade winds in both the great oceans of both hemispheres.

The following table of the northern limits, in different longitudes, of the northeast trade-wind in the Atlantic Ocean was given by Maury:[13]

Longitude W.	Latitude of Commencement of N.E. Trades in—			
	Winter.	Spring.	Summer.	Autumn.
70°	28°	28°.7	29°.3	29°.0
65	26.3	28 .0	29 .3	28 .3
60	24	24 .3	27 .3	28 .3
55	22	22 .7	24 .7	25 .0
50	21	23 .7	28 .3	23 .7
45	23	24 .7	31 .3	28 .7
40	27.7	29 .7	30 .7	29 .3
35	26	27 .3	30 .7	25 .7
30	24.3	28 .7	29 .7	26 .7
25	25.3	24 .7	31 .3	26 .3
20	24.3	28 .3	28 .7	27 .0
15	29	31 .0	32 .0	31 .3
10	....	31 .3	34 .7	32 .0

The following tables were prepared by Woeikoff from the "Pilot Chart of the Atlantic Ocean," edited by the Meteorological Office in London:[20]

MEAN POLAR LIMITS OF THE N.E. TRADE-WIND.

Months.	Meridians West.										
	65	60	55	50	45	40	35	30	25	20	17
Jan. to Mar..	26°	25°	23°.5	23°	24°.5	20°	26°.5	25°.5	25°.5	28°.5	30°
Apr. to June.	28.5	24.5	23	25	27	28	28	28	28 .5	32	33
July to Sept.	27	27	26 .5	26	26 .5	27.5	27 .5	28 .5	31	31 .5	32.5
Oct. to Dec..	26	24	22 .5	22	22 .5	24.5	25 .5	25 .5	26 .5	29 .5	31

MEAN POLAR LIMITS OF THE S.E. TRADE-WIND.

Months.	Meridians.									
	30 W.	25 W.	20 W.	15 W.	10 W.	5 W.	0	5 E.	10 E.	15 E.
Jan. to March.....	19°.5	21°	24°	26°.5	28°	29°	30°	31°.5	32°.5	33°
April to June......	25 .5	23	24	25	25	27	28.5	32	33 .5	
July to Sept.......	20 .5	22.5	24	24 .5	27.5	28.5	29.5	29 .5	30 .5	
Oct. to Dec.......	16 .5	18.5	20.5	21	22.5	28	28.5	29	30	

EQUATORIAL LIMITS OF THE NORTHERN AND SOUTHERN TRADE-WINDS.

Months.		Meridians West.					
		40	35	30	25	20	17
Jan.	N. E..........	3° N.	1°.5 N.	2° N.	4°.5 N.	6°.5 N.	8° N.
	S. E..........	1 N.	0 .5 N.	1 N.	2 N.	3 N.	3 N.
March	N. E..........	1.5 N.	0	0.5 N.	2 .5 N.	5 N.	6 N.
	S. E..........	1 S.	0 .5 S.	1 S.	0 .5 N.	0 .5 N.	1 N.
May	N. E..........	3.5 N.	3 N.	3.5 N.	5 .5 N.	8 .5 N.	
	S. E..........	0.5 S.	0	2 N.	3 N.	3 .5 N.	
July	N. E..........	8.5 N.	9 N.	10 N.	12 N.	14 N.	
	S. E..........	4 N.	4 N.	3 N.	3 N.	3 N.	
Sept.	N. E..........	11.5 N.	12 N.	11.5 N.	11 N.	12 N.	
	S. E..........	6 N.	4 N.	2 N.	2 N.	0 N.	
Nov.	N. E..........	6 N.	6 N.	6 N.	6 N.	9 .5 N.	
	S. E..........	4.5 N.	4 N.	3.5 N.	3 .5 N.	4 N.	

108. The first of the last three tables does not differ more from the preceding one, given by Maury, than is to be expected in determining a limit so ill-defined as that of the northern limit of the northeast trade-wind, which is likewise the southern limit of the northern tropical calm-belt. Either of these tables indicates that there is an annual oscillation of this limit with a range of about three degrees of latitude, and having its most northerly position in midsummer. It is seen, also, that this limit does not coincide with any parallel of latitude across the Atlantic Ocean, but it has a more northerly position on each side of the ocean than in the middle, the longitude of minimum latitude being that of about 50° W.

There are observations that indicate that the northern limit of this tropical calm-belt has a corresponding oscillation. When this belt in the fall moves southward from its most northerly position, it has been observed that the southwest winds are felt first on the coast of Portugal, then at Madeira, and afterwards at Teneriffe and the Canaries.

The southern limit of the southeast trade-wind is not so well determined from observation; but according to the second of the last three tables, this wind extends to about the same distance from the equator as the northeast trade-wind, and it has

also an annual oscillation corresponding to that of the northern limit of the northeast trade-wind, having its most northerly position about the same time, as required by theory, but this oscillation is not so marked, the range being smaller. In this case the limit lies nearest the equator on the west side of the ocean, and is about ten degrees further towards the south on the other side.

According to the last of the preceding tables, the equatorial limit of the northeast trade-wind, which is likewise the northern limit of the equatorial calm-belt, has a range of annual oscillation of about twelve degrees of latitude, from 20° to 40° W. longitude, with a position, at all seasons, a little more northerly on the east than on the west side of the field of observation, the extreme northerly position here in summer being about the parallel of 13° N. But the equatorial limit of the southeast trade-wind, which is also the southern limit of the equatorial calm-belt, has a much smaller range, for reasons given in § 102, it being only about five degrees on the west side on the longitude of 40° W., and sensibly vanishing on the east side of the ocean. The most southerly position, in March, is about one degree south of the equator.

Since the northern limit of the equatorial calm-belt has a much greater range of oscillation than the southern limit, especially on the east side, it follows that this calm-belt is much wider in summer than in winter, and especially so on the eastern side of the ocean.

109. Kerhallet, according to Dove,[13] gave the following table of the extent of the trade-winds in the Pacific Ocean, compiled from the observations of 92 ships:

The irregularity in these numbers indicates that they are not very accurate; but still they plainly show, as in the Atlantic Ocean, that the limits of both the northeast and southeast trade-winds have an oscillatory movement with a range, in the equatorial limits at least, of several degrees of latitude, and that the mean position of the equatorial calm-belt lies north of the equator, and has a greater width in summer than in winter. This greater width here, as in the Atlantic Ocean, it is seen

CONSIDÉRATIONS GÉNÉRALES SUR L'OCÉAN PACIFIQUE, 1856.

Month.	Polar Limit.		Equatorial Limit.		Breadth of Calm-belt.
	N.E. Trade, Lat. N.	S.E. Trade, Lat. S.	N.E. Trade, Lat. N.	S.E. Trade, Lat. N.	
January	21° 0′	33° 25′	6° 30′	5° 0′	3° 30
February	28 28	28 51	4 1	2 0	2 1
March	29 0	31 10	8 15	5 50	2 25
April	30 0	27 25	4 45	2 0	2 45
May	25 5	28 24	7 52	3 36	4 16
June	27 41	25 0	9 58	2 30	7 28
July	31 43	25 28	12 5	5 4	7 1
August	29 30	24 18	15 0	2 30	12 30
September	24 20	24 51	13 56	8 11	5 45
October	26 6	23 27	12 20	3 32	8 48
November	25 9	28 39	......	....	....
December	24 0	22 30	5 12	1 56	3 16

from the table, is due to the greater range of oscillation of the northern than of the southern limit of the belt.

We are not advised with regard to the part of the ocean where the observations were mostly made, so that it is uncertain whether these results are applicable to the belt generally in this ocean, and whether the mean positions of the limits and the range of oscillation may not differ considerably at different longitudes.

It had formerly been supposed that the mean position of these limits and of the central parts of the calm-belts in this ocean were nearly symmetrically situated with regard to the two hemispheres, the central line of the equatorial calm coinciding with the equator. Kaemtz places the limits of the northeast trade-wind at 23° and 2° N., and those of the southeast trade-wind at 21° and 2° S. But the equatorial calm-belt in this ocean, as in the Atlantic Ocean, doubtless lies a little north of the equator.

CHAPTER IV.

CLIMATIC INFLUENCES OF THE GENERAL CIRCULATION.

ON THE RELATIVE CLIMATES OF THE LOWER AND HIGHER LATITUDES.

110. In consequence of the interchange of equatorial and polar waters, especially in the southern hemisphere, as explained in § 68, the equatorial regions of the globe are colder, and the polar regions warmer, than they otherwise would be. A similar, but perhaps a much smaller, effect is produced by the interchange of the warm air of the equatorial, and the cold air of the polar, regions in the general circulation of the atmosphere. Although the volume of air interchanged in a given time is much greater than that of the water of the ocean, yet the mass is probably less, since the mass of the whole atmosphere is only equal to that of an ocean covering the whole surface of the earth of about 10 meters in depth, and as the specific heat of air is only 0.2375, the whole capacity of the atmosphere for heat is only equal to that of an ocean of 2.5 meters in depth. If we suppose the ocean to have a uniform depth of 5000 meters (3.1 miles), the interchanging velocity in the case of the atmosphere would have to be 2000 times as great as that of the ocean to give rise to an equal transfer of heat from the equatorial to the polar regions. The velocity of interchange in the case of the atmosphere is, of course, much the greater; but it can scarcely be regarded as being 2000 times as great as that of the ocean, for it must be remembered that the motion of the atmosphere, except at the earth's surface, and at all altitudes near the equator, is mostly easterly or westerly, the polar and equatorial components of motion being generally small in comparison with the east and west components. It is probable, therefore, that the amount of heat

transferred from the equatorial to the polar regions is less in the case of the atmosphere than in that of the ocean. But we have reason to think that the difference of temperature between the equator and the polls is very much diminished by the latter, and therefore the effect of the former in the same direction must be at least quite sensible.

The relation between the hygrometric states of the atmosphere of the lower and higher latitudes is also affected by the general vertical circulation. In the middle and higher latitudes there is a gradual settling down of the atmosphere toward the earth's surface, where the temperature is higher than it is above; and so here the relative humidity gradually becomes less, and being very small above it becomes still much less after having descended to the lower strata. In the lower latitudes, and especially near the equator, on the contrary, where there is a gradual ascent of the air from the lower strata, where the temperature is much greater, to the upper ones, where it is much less, the relative humidity is gradually increased, even, it may be, to complete saturation. Of course the absolute amount of vapor in the higher latitudes, on account of the lowness of temperature, is much less than in the lower latitudes, but if there was the same relative humidity in both, the effect of the vertical circulation would be to decrease that of the higher latitudes and to increase that of the lower latitudes.

WET AND DRY ZONES.

111. As all the vapor of evaporation, or nearly, over both trade-wind zones of about 20° in width, especially where the trade-winds are regular, as on the ocean, is carried into the comparatively narrow zone of the equatorial calm-belt before it ascends to sufficient height to be condensed into rain, the amount of rain falling in this belt is very large and the cloud nearly continuous. The equatorial calm-belt, therefore, is also a cloud- and rain-belt. The height to which the vapor of the surface currents ascends before condensation and cloud-formation take place depends upon the amount of moisture in the

air at the earth's surface, and so upon the difference between the air temperature and that of the dew-point. The height at which condensation commences is given accurately by Table IV, and is nearly, under all conditions, 125 meters for each degree Centigrade of the depression of the dew-point below the air temperature, as explained in § 27.

The first vapor condensed is supported and carried farther up by the ascending current in the form of cloud and small particles of rain, and the more the stronger the current is. Small particles are carried upward by the current, and the still smaller ones more rapidly than the larger ones, and so the different sizes of the smaller drops, being carried up with different velocities, come in contact and combine until they form drops too large to be supported by the ascending current, and these then fall as rain. The stronger the ascending current the larger the drops. When there is no ascending current there is no rain, though there may be fog and mist.

The daily amount of evaporation on the ocean within the tropics is about one fourth of an inch per day. If then all this amount of vapor over zones say 1000 miles in width on each side is carried into the calm-belt say 300 miles in width, and is there condensed and falls as rain, then the daily rainfall is 1.67 inches per day; and if this belt were to remain stationary, it would be, for the average of the width, about 60 feet per year. But since the cloud- and rain-belt, as the calm-belt, oscillates through a range generally more than twice as great as its width, this amount of rain is distributed, in the course of the year, over a zone more than three times as wide, and hence, in general, less than one third of this amount falls in any one place during the year.

The calm-belt, as it exists at any given time, is mostly narrower than the belt on which rain falls; for as the air in the equatorial region generally, but mostly over the surface calm-belt, ascends it gradually expands and diverges from the central line of this belt to form the upper return current toward the poles; and before the vapor is all condensed, and while the air is still ascending, it is carried out beyond the limits of the

calm-belt, so that considerable rain falls at some distance beyond these limits.

Of course neither the calm-belt nor the rain- and cloud-belt has very definite limits, but these are much better defined over the great oceans, where the trade-winds blow very steadily in nearly the same direction and with the same force, than on the continents, where regularity of outline is very much interfered with by the various abnormal disturbances of uneven surfaces and mountain ranges, and likewise by the monsoons of the Indian Ocean and others. The rain-belt is, however, clearly traceable across the whole of Africa, wherever observations have been made, as also across the American isthmus, but it has greater width, and its limits are not so well defined. The zone on which rain falls during the course of the year has been given on a chart by Woeikoff.[11] This, on the great oceans, lies mostly between 1° and 11° north latitude; but on the west coast of Africa it extends a little farther north, and on the east coast of South America to about 5° south latitude. These limits correspond pretty closely with the extreme northern limit of the southeast trade-wind in January on the one hand, and the extreme southern limit of the northeast trade-wind in July on the other, as given in the tables of § 107 and § 109; and so the rainy zone falls nearly within the range of oscillation of the calm-belt.

112. Loomis[11] has given a chart of mean annual rainfall on which the northern limit of the zone of 50 inches and upward of mean annual rainfall, on the continent of Africa, lies mostly between the parallels of 10° and 12° north latitude, but the southern limit is more irregular and extends from a point near the equator on the west coast to about 10° south on the east coast of Africa. Beyond these limits, both north and south, there is a gradual shading off on the chart and diminution in the amount of rainfall, so that although the rainy zone here has no definite limits, yet the amount of rain falling a little north of the equator, on the parallels of latitude corresponding to those of the rain-belt on the oceans, as the result of the general circulation of the atmosphere and meeting near the equa-

tor of the vapor-laden surface currents of the trade-winds is unusually great, notwithstanding the amount of evaporation over the land surface is much less than on the ocean, and this vapor, on account of the inequalities of the land surface and the frictional resistances, is not all carried so nearly into the central line of the meeting of the trade-winds as on the smooth ocean surface.

Farther east, over the islands of Sumatra, Java, Borneo, and Celebes, and also over the Malayan peninsula, all lying on and near the equator, the mean annual rainfall, according to the chart, is still more abundant; for they all fall within the zone of 75 inches and over of rainfall, and at many individual stations the annual rainfall is over 150 inches, and even more than 200 inches at Buitenzorg, Java, all going to show the tendency to a concentration of rainfall in a zone at and near the equator. The amount of rainfall here is much greater than over the continent of Africa, because the amount of evaporation over the warm ocean surface is greater; but on account of the monsoon influences, to be considered hereafter, this rainfall is scattered over a much wider zone than on the ocean generally, and consequently the rainfall in the central parts of the rainy zone is less, and the zone has no definite limits.

In the northern part of South America, a little north of the equator, this chart does not indicate that there is a distinct and well-defined rainy zone, as on the same parallels on the two great oceans, because it is mostly obscured by the rainfalls over a great part of South and Central America, arising from the influence of the mountain chain of the Andes, to be explained farther on. There is much evidence, however, to show that here also there is, at least in many places, a distinct rain-belt, oscillating with the seasons, and having, at any given time, definite limits beyond which little or no rain falls.

The latent heat given out in the condensation of the great amount of aqueous vapor carried into, and ascending in, the calm-belt gives additional strength to the trade-winds. In consequence of this heat the upper part of the air, above the plane of incipient condensation, is not cooled by expansion down to

the temperature of the air on each side, and hence the tendency to ascend in this belt is very much stronger than that of the air on each side, which simply ascends with sufficient rapidity to supply the returning polar currents above, in the general vertical interchange between the equatorial and polar regions, which, we have seen, is small, the force arising from the difference of temperature between the equator and the poles being almost entirely counteracted by the deflecting forces arising from the earth's rotation. Hence there is a very rapid ascent of air in this belt, which requires an increased equatorial component of velocity in the trade-winds near the earth's surface to supply it, and this adds greatly to their strength. This is entirely a secondary matter, dependent upon the general vertical circulation, for if it were not for this the air, with its aqueous vapor, would not ascend at all, and hence no condensation would take place.

113. The air within the rain-belt being almost entirely calm, and also warm and nearly saturated with vapor, is always extremely oppressive, such as is sometimes experienced for a short time in other regions of the globe when the air is warm and calm, and saturated, or nearly so, with aqueous vapor. A graphic description of the kind of weather which is usually experienced under the cloud-ring of the equatorial calm-belt is found, as cited by Maury,[21] in the journal of Commodore Sinclair, kept on board of the United States frigate Congress during a cruise to South America in 1817–18. He crossed it in the month of January, 1818, between the parallel of 4° north and the equator, and between the meridians of 19° and 23° west. He says of it:

"This is certainly one of the most unpleasant regions on our globe. A dense, close atmosphere, except for a few hours after a thunder-storm, during which time torrents of rain fall, when the air becomes a little refreshed; but a hot, glowing sun soon heats it again, and but for your awnings, and the little air put in circulation by the continual flapping of the ship's sails, it would be almost insufferable. No person who has not crossed the region can form an adequate idea of its unpleasant effects. You feel a degree of lassitude unconquerable, which not even sea-bathing, which everywhere else proves so salutary and renovating, can dispel.

Except when in actual danger of shipwreck, I never spent twelve more disagreeable days in the professional part of my life than in these calm latitudes.

"I crossed the line on the 17th of January, at eight A.M., in longitude 21° 20', and soon found I had surmounted all the difficulties consequent to that event; that the breeze continued to freshen and draw around to the south-southeast, bringing with it a clear sky and most heavenly temperature, renovating and refreshing beyond description. Nothing was now to be seen but cheerful countenances, exchanging as by enchantment from that sleepy sluggishness which had borne us all down for the last two weeks."

But notwithstanding the oppressive character of the weather under the cloud-ring the temperature is not extremely high—much less so than on either side of it, and the disagreeable and oppressive character arises mostly from the calmness and extreme dampness of the air. The great cloud-ring protects the earth's surface from the almost vertical rays of the sun, and hence it is considerably cooler than on either side where the sun's heat reaches the earth's surface unobstructed. Espy found from examining the tables of Wilkes that in the rainy belt of the equator the temperature of the air on the ocean is about 6° F. lower than it is beyond the borders of the rains both north and south. On land the difference is much greater. According to Humboldt, each side of the belt of rains in South America is from 10° to 18° hotter than in the belt of rains. This is caused, not only by the absence of the heating rays of the sun in the rain-belt, but also by the cooling effect of the drops of rain coming down from high and cold altitudes.

114. From the general circulation of the atmosphere and meeting of the trade-winds near the equator, and the annual oscillation and shifting of the parallels on which they meet, arise peculiarities of climate within the zone of the range of oscillation of the calm- and rain-belt, such as are not to be found elsewhere. Since the calm-belt with its daily torrents of rain oscillates forth and back annually through a range, especially on the great oceans, more than twice as great as its width, while just outside of the limits of this narrow belt, with winds from different directions on the two sides, there is no rainfall at the

time, this is necessarily a zone of alternations of extremely wet and extremely dry seasons, and of winds which vary in direction at different seasons of the year. The position of a place with regard to the mean position of the oscillating rain-belt may be such that there is either a long rainy and a long dry season during the year, or two rainy seasons of shorter duration with two intervening dry seasons. The latter takes place where the station is situated at or near the mean position of the middle of the rain-belt, and the range of oscillation is considerably greater than the width of the belt, as it generally is, on the oceans; for then the middle of the belt, as it moves towards the north, is over the place of observation in April or May, and again in October and November as it moves back southward, and at these times there is a great abundance of rain. But in July or August, when the rain-belt has its most northerly position and its southern limit is north of the place, or in January or February, when it has its most southerly position, and its northern limit is south of the place, there is of course no rainfall. There are, consequently, in this case, two wet and two very dry seasons in the course of the year. Or the width of the rain-belt and the range of oscillation, especially on land, may be such that there is no total cessation of rain during the year, but simply two minima instead of the two dry seasons, and so two maxima, one in the middle of each of the rainy seasons.

If the place of observation is a little north of the mean position of the rain-belt, so that its southern limit never moves north of the place, then there is no intervening dry season when the belt has its extreme northerly position, though there may be a considerable diminution in the rate of rainfall in August or September, when a minimum only occurs instead of a complete cessation and a very dry season, but a long drought at the opposite season of the year. On the other hand, if the place is so far south of the mean position of the oscillating rain-belt that the northern limit of the belt, in its extreme southerly position, does not lie beyond it, then there is no complete cessation of rain in January or February, but simply a minimum

in the monthly rainfalls, preceded and followed by a maximum, while at the opposite season of the year there may be no rain for several months. The position of the place also and the width of the rain-belt may be such as to cause a rainy season without a minimum in the middle, and in such cases the dry season is much longer than the rainy season.

Since on the north side of the rain-belt the winds are generally from some point between north and east, and in the other from some point between south and east, wherever the rain-belt, in its annual oscillations, passes entirely over a place, there is a change of the winds from some point in the northeast to one in the southeast quadrant, or the contrary, Since the calm-belt is sometimes considerably north of the equator, the southeast trade-winds, after passing the equator, in consequence of the influence of the earth's rotation, then become southerly or even southwesterly winds near the southern border of the calm-belt.

115. The preceding deductions from theoretical considerations are verified by numerous observations of the monthly rainfalls and prevailing directions of the wind at many places near the equator, all around the globe. Since the oscillating calm, and the accompanying rain-belt always follow each other and occupy the same parallels, except that the latter is a little the wider, the oscillations of the calm-belt can be more readily traced from the observations of the monthly rainfalls than from the direct observations of the winds and calms.

The following table of average monthly rainfalls on the Congo and the southwest coast of Africa has been taken from a work by Dr. A. v. Danckelman.[24]

At Vivi, 5° 40′ S., 13° 49′ E., it is seen that there is a total cessation of rain during June, July, and August, and very nearly so in September, while at the opposite season of the year there is only a minimum with a maximum preceding in November and one following in April. The station being south of the equator, during the summer of the northern hemisphere, when the rain-belt has its most northerly position, the southern border of this belt was entirely north of the place, notwithstanding it is wider at this land station than on the ocean and its

Month.	Monthly Rainfalls in Millimeters at				
	Vivi.	Ponta da Lenha.	Gabun.	Chinchoxo.	Loanda.
January	96	65	155	311	39
February	68	104	217	120	30
March	103	94	349	185	58
April	231	118	381	102	122
May	50	27	169	54	12
June	0	0	7	0	0
July	0	0	1	0	0
August	0	0	11	5	1
September	1	1	40	8	2
October	75	3	288	23	4
November	237	143	423	222	51
December	180	53	224	52	25
Year	1,041	608	2,265	1,082	344

limits not so well defined, and so no rain fell for several months. The minimum in February, if there are no abnormal local disturbances, indicates the middle of the rain-belt at this time is south of the station, and if so, it must have a more southerly mean position or a wider range than on the Atlantic Ocean. What has been stated here with regard to Vivi is also true of Ponta da Lenha.

At Gabun, 0° 30′ N., 9° 35′ E., since it lies farther north, the rainfall does not quite vanish in the summer of the northern hemisphere, and the minimum of the opposite season is less marked than at Vivi. The same is true with renard to Chinchoxo, except that here there seems to be a total cessation of rain during June and July, but the numbers expressing the monthly rainfalls seem to be very irregular, and so are perhaps not very reliable.

At Loanda, 8° 49′ S., 13° 7′ E., there is much less rain than at Vivi; but there are two maxima and two minima, with an intervening minimum in the winter of the northern hemisphere and a total cessation of rain for several months at the opposite season of the year, as at Vivi.

At the island of St. Thomas, 0° 20′ N., 6° 43′ E., and so of nearly the same latitude as Gabun, from the average of five or

six years of observation, no rain fell in July; in June, July, August, and September only 47 mm. of rain fell; while the amount for the whole year on the average was 1019 mm. The greatest amount fell in March.[35]

116. "In Nango, 13° N., 9° W., from Paris, the rainy season commences the latter part of May, but the rain is then only seldom. It becomes more frequent in July. The rainfall then is very great. The temperature then sinks still more. The rain holds on in August, and is as abundant as in July. The temperature is then a minimum. September brings a great change, but the rain is still frequent. The temperature rises again, and the air becomes more clear."[36]

This place, being 9° north of the equator, is within the rain-belt from June to September inclusive, and the rainfall during the middle part of this period is abundant; but during the other eight months of the year, when the rain-belt has moved southward, the northern limit of the belt is south of the place, and hence there is no rain.

Observations on the upper Nile in the interior of Africa, both of the winds and the rainfall, indicate clearly the existence of the calm- and rain-belt there. At Rubaga, 0° 20′ N., 32° 45′ E., altitude 1300 meters, the mean probability of rain is as follows:

Jan.	Feb.	Mar.	April.	May.	June.	July	Aug.	Sept.	Oct.	Nov.	Dec.
43	42	42	70	51	60	29	46	52	74	67	47

There are, therefore, two maxima and two minima, as usual at places near the equator; but the numbers indicate that the rain-belt here is wide and not well defined, and so, instead of two rainy and two dry seasons, there are simply two maxima and two minima of rainfall.

At Lado, 5° 2′ N., 31° 50′ E., the winds from April to September are mostly from S.W. and S.; from October to March, N. and N.E. The greatest rainfall is in August and September; the least in December, January, and February.[38]

These observations, as well as those of Rubaga, indicate that the calm- and rain-belt here is but little north of the equator, and that its width and range of oscillation are such that the places are at all times comprised between its limits;

but since there is more rain during the summer than the winter of the northern hemisphere, the mean position of the middle of the belt must be south of Lado. The winds, however, indicate that the mean position cannot be far from Lado.

"At Nyassa Sea (south of the equator) there is no rain between May and October. The soil becomes dried out, and the grass becomes dry and assumes the appearance of wheat-fields in England in July."

During this dry season the rain-belt is north of the equator, and Nyassa Sea is left in the zone of the southeast trade-winds, where there is no rain.

During Emin Effendi's travels in equatorial Africa, from Nov. 29 to Dec. 18, 1877, from Mréili, 3° N., to Rubaga (Mtesa's residence), rain fell nearly every day in great abundance, and the whole country was flooded with water, and the travellers had to wade through water often from two to three feet deep."

The middle line of the rain-belt, therefore, at this time, which is about the time of its mean position, must have been about two degrees north of the equator.

117. In the northern part of South America and the Isthmus of Panama the middle line of the rain-belt seems to oscillate between the equator and the parallel of 10° or 12° N,; and so, in its mean position, is considerably north of the equator.

At Guatemala, 14° 38′ N., 90° 31′ W.; altitude 1480 m., from three years of observations, 1880–1882, the average monthly rainfall in millimeters is:

Jan.	Feb.	Mar.	April.	May.	June.	July.	Aug.	Sept.	Oct.	Nov.	Dec.	Year.
4	6	1	20	144	253	137	234	231	159	51	2	1242

Hence this place is fully within the limits of the rain-belt from May to November, but during the rest of the year, while the belt has a more southerly position, it is almost wholly north of the northern limit of the belt, where it receives little or no rain. There is an indication of a slight minimum in July also, from which it seems that the middle of the rain-belt passes a little north of Guatemala at this time.

At Greytown, Nicaragua, lat. 10° 55′ N., according to Commander A. V. Reed, "the dry season lasts till late in May, when the rains set in and are of daily occurrence during June and July, with considerable thunder and lightning. In August and September, they have very pleasant weather, with less rain, and the sea outside is usually smooth; but in October and November again they have disagreeable weather and daily rains."

Here again, it seems, there are two maxima, one in June or July and the other in October or November, with a minimum in August or September, and a long dry season at the opposite season of the year. This indicates that the mean position of the middle of the rain-belt is considerably south of Greytown; for at any place in the middle of the belt in its mean position the two maxima would occur at exactly opposite times of the year, and there would be two intervening dry seasons of equal length, in which there would probably be one or two months with scarcely any rain.

From the report of Lieut. Frederick Collins of the survey for an Interoceanic Canal in the State of Cauca, we learn that, at the junction of the Napipi and Merindo rivers (about 5° N.), "as a rule, two well-marked dry seasons are experienced here, with corresponding periods of rain. January, February, and March are the months which constitute the dryest and pleasantest season. In April the rains commence, and in May and June they are very heavy. In July a second dry season begins to set in, and August and September are generally pleasant and comparatively dry. In October rains again commence, and in November and December they are the heaviest."

Here the mean position of the middle of the rain-belt seems to be a little south of the place, since the dry season of August and September is not so marked and of as long duration as that of January, February, and March.

On the Panama Interoceanic Ship Canal, 8°.5 N., "the year is divided, as in other parts of the American isthmus and of Central America, into two seasons, the rainy and the dry, the former beginning in the latter part of May and lasting until November, when it gives place to the latter, which lasts until

May comes around again. The annual rainfall is from 90 to 140 inches."[28]

This place being a little farther north than the preceding one, there are here only a wet and a dry season, the former taking place while the belt is north of its mean position; but, as in other places considerably north of the equator, if we had the monthly rainfalls, a minimum in August or September would undoubtedly be observable.

The effect of the oscillation of the rain-belt near the equator in producing a wet and a dry season seems to be felt in Mexico. Humboldt says:

"There are only two seasons known in the equatorial region of Mexico, even as far as the 28th degree of N. latitude: the rainy season, *estacion de las aguas*, which begins in the month of June or July, and ends in the month of September or October; and the dry season, *el estio*, which lasts eight months, from October to the end of May."[29]

118. Although the rain falls every day in the rain-belt during at least the middle part of the rainy season, yet it is not continuous through both the day and night, but it takes place mostly during the day. The abundant downpour of rain, at least on land, seems to be of the nature of thunder showers in all countries during very warm and damp weather, and to depend upon the heating effects of the sun's rays in inducing a state of unstable equilibrium, so that the aqueous vapor brought in by the trade-winds into the calm-belt ascend and are condensed into rain, mostly while the atmosphere is in this state. Tomlinson says:[30]

"The atmospheric phenomena of such countries as lie under the calm-belt are marked by a striking uniformity and regularity. For hours after the rising of the sun the sky remains entirely free from clouds. About noon a few clouds appear at the horizon, which, as they become more vertical, rapidly increase in size and density; presently the thunder is heard, the wind blows in violent gusts, and at the same time there occurs a heavy downpour of rain, which lasts for a few hours. The clouds quickly disappear toward evening, and the sun sets in a deep-blue and cloudless sky. It is rare for rain to fall during more than eight hours at a time. The countries near the equator experience this kind of weather nearly all the year, and what is called the rainy season differs from the

remainder of the year in the rain being rather more continuous and plentiful."

This, or at least the latter part, seems to apply more especially to continents where the rainfall is more scattering and not well defined, and not to the ocean, or even to the American isthmus, where, we have seen, there are long seasons with little or no rain.

Since the origination of the shower or daily rainfall depends upon the unstable state, which is more readily induced during the heat of the day, the rain commences at that time; but after an abundant fall of rain for several hours, some of which descends from very high altitudes where the air and the condensed vapor are very cold, the unstable state of the atmosphere toward evening is destroyed by the cooling effect of the rain upon the lower strata of the atmosphere, and the rain ceases until next day, when the unstable state is again induced.

119. The following contrasts of the wet and the dry seasons on the Orinoco and the great Amazonian basin within the range of oscillation of the rain-belt, and which is true of all places within this range where the rain-belt is narrow and well defined, and not much affected by the abnormal disturbances, but is somewhat as it is on the ocean and on level countries near the ocean, have been given, as cited by Maury,[11] by Humboldt in his "Aspects of Nature:"

"When, under the vertical rays of the never-clouded sun, the carbonized turfy covering falls into dust, the indurated soil cracks asunder as if from the shock of an earthquake. If at such times two opposing currents of air, whose conflict produces a rotary motion, come in contact with the soil, the plain assumes a strange and singular aspect. Like conical-shaped clouds, the points of which descend to the earth, the sand rises through the rarefied air on the electrically charged centre of the whirling current, resembling the loud water-spout, dreaded by the experienced mariner. The lowering sky sheds a dim, almost straw-colored light on the desolate plain. The horizon draws suddenly nearer, the steppe seems to contract, and with it the heart of the wanderer. The hot, dusty particles which fill the air increase its suffocating heat, and the east wind, blowing over the long-heated soil, brings with it no refreshment, but rather a still more burning glow. The pools, which the

yellow, fading branches of the fan-palm had protected from evaporation, now gradually disappear. As in the icy north the animals become torpid with cold, so here, under the influence of the parching drought, the crocodile and the boa become motionless and fall asleep, deeply buried in the mud. . . .

"The distant palm-bush, apparently raised by the influence of the contact of unequally heated and therefore unequally dense strata of air, hovers above the ground, from which it is separated by a narrow intervening margin. Half concealed by the dense clouds of dust, restless with the pain of thirst and hunger, the horses and cattle roam around, the cattle lowing dismally, and the horses stretching out their long necks, and snuffing the wind, if haply a moister current may betray the neighborhood of a not wholly dried-up pool. . . .

"At length, after the long drought, the welcome season of the rain arrives; and then how suddenly is the scene changed! . . .

"Hardly has the surface of the earth received the refreshing moisture when the previously barren steppe begins to exhale sweet odors, and to clothe itself with killingias, the many panicles of the paspulum, and a variety of grasses. The herbaceous mimosas, with renewed sensibility to the influence of light, unfold their drooping, slumbering leaves to greet the rising sun; and the early song of birds and the opening blossoms of the water plants join to salute the morning."

120. We have seen that under the high pressure of the tropical calm-belts there is a gradual settling down of the air to supply the outward flow on each side at the earth's surface from beneath this high pressure. The aqueous vapor in these belts cannot rise up so as to be condensed, but, on the contrary, it rather sinks down and is carried away by the surface currents; on the one hand, by the trade-winds to the equatorial calm-belt, and on the other, away toward the middle latitudes. The tropical calm-belts are therefore dry belts. Over the trade-wind latitudes, also, since the horizontal and comparatively strong surface currents hurry away the vapor which is evaporated over the ocean and the land surface of these latitudes to the equatorial calm-belt, where alone it rises up to where it can be condensed, there is scarcely any rainfall. There is, therefore, in each hemisphere, a wide zone, extending from the zone of equatorial variable rains, over the trade-wind latitudes and comprising the tropical calm-belt, in which there is little rain, especially on the oceans, where the general vertical

circulation is more regular. It is true there is considerable rain in some places within these zones, but this is due to abnormal disturbances of mountain ranges, and not to the general circulation of the atmosphere, such as would exist if the whole surface of the earth were homogeneous.

These zones are almost completely rainless on most parts of the great oceans, but there is at least a strong tendency toward this same condition on the continents, as is readily seen from an inspection of Loomis's chart of Mean Annual Rainfall. In the northern hemisphere between the parallels of 15° and 40°, are the dry regions of California, Arizona, and Colorado in North America, the great Sahara and Nubian deserts of North Africa, and the dry region of Arabia and Persia in Asia. In the southern hemisphere within the same parallels are the dry regions of the southern part of the Argentine Republic and of eastern Patagonia in South America, a large dry region in South Africa, and one comprising the whole of the interior of Australia. It must not be understood that these regions are entirely rainless throughout the year, but simply that the amount of rain is very scant, and small in comparison with that of the middle latitudes of each hemisphere. Even in the great Sahara rain sometimes falls.

RELATIVE TEMPERATURES OF EAST AND WEST SIDES OF THE CONTINENTS.

121. One of the principal climatic effects arising from the general circulation of the atmosphere is the great difference of temperatures observed on the same latitudes on the east and west sides of the continents, especially in high latitudes, both in the annual means of temperature and also in the extreme temperatures of the seasons. It is a matter of observation, which has been explained in § 68, that the mean annual temperatures of the oceans in the higher latitudes are greater, and those of the lower latitudes less, than those of the same latitudes on the continents; and the differences would be still greater if it were not for the equalizing tendency of the general

circulation around the globe, easterly in the higher, and the contrary at the earth's surface in the lower, latitudes. The effect, also, of this circulation is to cause the highest mean annual temperatures of the oceans in the higher latitudes to be on the east sides, near the continents, and the lowest mean temperatures of the continents to be on the east sides of the continents near to the adjacent oceans. Hence there are the greatest contrasts between the east sides of the oceans and of the continents. From a chart of isotherms for the mean temperatures of the year it is seen that the mean annual temperature of the Atlantic Ocean near the coast of Norway is 40° F., while on the same latitude in the eastern part of Siberia it is only about 10° F. There is also a difference of about 20° between the mean annual temperatures of the British Isles and those of the same latitudes on the eastern side of Asia. Likewise the difference of mean annual temperatures between the North Pacific Ocean, adjacent to Alaska in the neighborhood of the Aleutian Islands, and those of corresponding latitudes of Hudson's Bay and Labrador, is about 15°, but for latitudes a little lower, some less.

Since the annual inequalities of temperature are very great on the continents in comparison with those on the ocean, the greatest contrasts of temperature are found in the winter season. For since the mean annual temperatures on the ocean in the higher latitudes are considerably above those of the same latitude on the continent, and the range of annual oscillation is small on the ocean, the mean midwinter temperatures scarcely fall to the mean annual temperatures of the land; and so the differences of the temperatures on the ocean and on the continent, when the east sides are compared, are equal to the whole differences between the annual mean and the midwinter temperatures of the latter. Thus the mean temperature of the ocean in January between Iceland and Norway, according to Buchan's charts, is about 35° F., while that of eastern Siberia is −40°, a difference of about 75°. Between the mean January temperatures of the ocean west of the British Isles and those of the same latitudes in the eastern part of Asia the difference

is less, but still about 60°. Between the eastern sides of the North Pacific and the corresponding latitudes on the east side of America the differences in January are nearly as great.

In midsummer the contrasts in the higher latitudes made in a similar manner are much less and reversed, the oceans at this season being colder than the continents. But still there are considerable differences when we compare the eastern sides of the oceans with those of the continents. The July mean temperature midway between Iceland and Norway is about 10° less than on the same latitude in the eastern part of Asia, and the difference between the Aleutian Islands and Labrador on the same latitude is about the same.

On the lower latitudes, those of the trade-winds, the differences between the mean annual temperatures of the continents and the oceans are considerable, those of the continents being the higher. The differences in midsummer are still greater, but in midwinter the temperatures are somewhat the same.

122. The characteristics of a continental climate in the middle and higher latitudes are great extremes in the annual changes of temperature with the maxima and minima occurring earlier in the season, a dry atmosphere, and a lower mean annual temperature than the normal of the latitude; while on the ocean the range of annual oscillation is very small, the maxima and minima occur later in the season, and the mean annual temperature is higher than on the continents on the same latitudes. Now it is readily seen that the effect of the general circulation from west to east around the globe in the middle and higher latitudes is not only to throw the true continental and oceanic climates from the middle over toward the eastern sides of the continents and oceans, but also to cause the western sides of the oceans to have a somewhat continental climate and the western sides of the continents to have in some measure an oceanic climate, and consequently to cause a great contrast of climates between the western and eastern sides of the continents. Accordingly Europe, especially the western side, has a mean temperature a little above the normal of latitude, small annual inequalities with the maxima and minima

occurring later in the season, and a damper atmosphere; while on the eastern side of Asia, on the same latitudes, the mean annual temperature is a little below the normal of latitude, and there are large annual inequalities of temperature with maxima and minima occurring earlier in the season, and a much drier atmosphere.

The mean annual temperatures of Norway and Sweden, from this cause, are about 20° F. higher than those of the eastern part of Siberia, and those of France about 15° higher than those of the Chinese Empire on the same latitudes. Similar differences of mean temperature are also found between the western and eastern sides of North America. Again, while the mean annual range of temperature in Norway and Sweden is scarcely as much as 35° F., that of the eastern part of Asia on the same latitude is more than 90°. The annual temperature ranges, also, in Germany, are only about half as great as on the same parallels of latitude in the Chinese Empire. Similar differences are observed in the annual inequalities of temperature between the west and east sides of North America, but not nearly so great, because the continent is much narrower. The epochs, also, of midwinter and midsummer occur about a half month later on the west than on the east sides of the continents, the former corresponding more nearly with those of oceanic, and the latter with those of continental, climates.

In the southern hemisphere the continents do not extend so far into the higher latitudes, and the contrasts of climate between the two sides of the continents are comparatively small, the tendency of the general circulation around the globe being to reduce the temperatures and all the climatic features somewhat to those of the ocean, so that in the southern parts of South America and of Africa there are only small annual temperature ranges.

In the lower latitudes of both hemispheres the annual temperature ranges of course are small, and consequently there is but little difference in temperature between the two sides of the continents, and the principal differences here are in the mean annual temperatures, those of the ocean being consider-

ably lower, but in consequence here of the westerly motion of the atmosphere at the earth's surface, the eastern sides of the oceans and the western sides of the continents have higher mean annual temperatures than the western sides of the oceans and the eastern sides of the continents. For instance, the easterly winds of the great Sahara, and North Africa generally, must tend to drive the heat and the dryness westward toward the Atlantic, so that the east side of the Atlantic here which receives the sand and dust from the continent, § 87, must have a higher air temperature and a drier air than the west side.

IN CONNECTION WITH MOUNTAIN RANGES.

128. If the whole surface of the earth were that of the ocean, or any smooth homogeneous surface, the calm-belts, the rain-belt, and the dry zones would extend without interruption entirely around the globe with the same regularity which is observed upon the oceans, and everywhere the same climatic conditions would exist on the same parallels of latitude. But on account of the influence of mountain ranges in deflecting the currents of the general circulation of the atmosphere, great diversities of climate are found in different places on the same parallels arising from this cause. The air of the lower strata of the atmosphere in the trade-wind zone of the North Atlantic, having a westerly motion, and impinging against the high table-lands and mountain ranges of Mexico, is deflected around toward the north over the southeastern States and up the Mississippi Valley into the higher latitudes, where it combines with the general easterly flow of these latitudes, and adds to its strength. This completely breaks up the continuity of the tropical calm-belt and dry zone, so that instead of a dry region with scanty rainfall, such as is found in North Africa, Arabia, Persia, Beloochistan, and Cabul, we have on the same parallels in the southern and eastern United States a region of abundant rainfall, and all the way up the Mississippi Valley and in the interior of the continent there is much more rain than in the interior of Asia. In consequence of this deflection, the trade-

winds of the Atlantic become nearly east winds in the vicinity of the West India Islands, and farther on in the southeastern States and the Mississippi Valley the winds are southeasterly and then southerly, and the warm and moist air of the Gulf of Mexico, being carried around into higher and cooler latitudes, not only furnishes vapor for condensation into rain, but also modifies somewhat the temperatures. There is, therefore, a sufficiency of rain here on the parallels of the north tropical dry zone for agricultural purposes as far as the 100th meridian west of Greenwich.

The easterly current across the Atlantic Ocean on the middle and higher parallels, strengthened a little by the deflected current coming from the Gulf up the Mississippi Valley, impinging against the inequalities of the coast of Europe, and especially the interior mountain ranges, is partly deflected both to the right and to the left, the one branch passing down toward the Canaries and the northwest coast of Africa, joins the general westerly flow in the trade-wind latitudes, while the other curves around along the coast of Norway on to Spitzbergen and around to Greenland.

In consequence of these deflections there is a gyration of air around a central region midway between America and the eastern continent on the parallels of the zone of high pressure, and also one around a central area in the vicinity of Iceland. These gyrations are indicated by the arrows showing the prevailing directions of the winds on Coffin's charts,[11] and also by the charts of M. Brault, based upon many thousands of observations. From the former of these gyrations it results that the northeast trade-winds adjacent to Africa seem to come from a more northerly direction and from a higher latitude than they do farther west, while on the west side of the ocean they become easterly, and even southeasterly, winds. The former also gives rise to the prevailing northwest winds in the part of the Atlantic Ocean adjacent to Spain and Portugal and the northwest coast of Africa, and the latter in part to the prevailing southwest winds of the British Isles and the coast of Norway, and the northeasterly winds of Greenland.

Where the general easterly current of the middle latitudes impinges against the high lands of the interior of Asia, comprising the Kuenlun and Himalaya chains, the latter 1300 miles in length, with an average height of 18,000 feet, on the south, and the Great and Little Altai Mountains on the north, with the elevated and desert table-lands containing the Desert of Gobi between them, they are again deflected, not around and back, as on the west coast of Europe, but to each side, the southern branch giving rise to prevailing northwesterly winds over a large region, including southeast Europe and southwest Russia. Woeikoff says: "There is a region comprising Hungary, Transylvania, the Danubian Principalities, and southwest Russia, in which the prevailing winds are northwest both winter and summer." [20] These winds also prevail over Arabia and Persia, and this deflection likewise affects the directions of the regular monsoons of India, and its effect is felt even at the equator south of Asia, where it gives rise to a strong west wind, sometimes called the *west monsoon of the line*, where, otherwise, there would be no wind with an east or west component of motion.

124. In like manner, in the lower latitudes of the South Atlantic Ocean the general westerly current of the lower part of the atmosphere, on arriving at the continent of South America, and especially at the high range of the Andes, with a mean elevation of nearly 12,000 feet, is deflected around toward the south over Brazil and on southeastward until it enters and combines with the strong easterly current of the middle latitudes of the South Atlantic. Instead, therefore, of the rainless zone, as it exists on the ocean in the trade-wind and high-pressure latitudes, extending across the continent of South America, or there being a deficiency of rain, as in some other countries on these latitudes, there is a sufficiency of rain over all Brazil and the eastern side of South America, up nearly to the parallel of 40° S.

The moisture for this rain is furnished by the deflected currents of warm and damp air from the trade-wind and rainy zone, for the air of the trade-winds is damp, but the zones of trade-

winds on the oceans are rainless, because the conditions, which are found on the uneven land surface, for producing ascending currents and vapor condensation are wanting on the oceans. On account of this deflection, instead of southeasterly winds, as in the ocean on the same latitudes, the prevailing winds are northeasterly and northerly over Brazil, just as in the southern part of the United States and the Mississippi Valley they are southeasterly and southerly, and the continuity of the tropical calm-belt and rainless zone of this hemisphere is also interfered with.

A small part also of the strong easterly current of the middle latitudes is deflected around toward the equator by the west coast and the mountain ranges of South Africa, so that there is, on the South Atlantic, as on the North Atlantic, a gyration of air around a central area, midway between South America and Africa, and on the parallels of the zone of high pressure. This is also plainly indicated by the charts of both Coffin and M. Brault. As a result of this gyration the southeast tradewinds come apparently from a higher latitude and have a more southerly direction on the part of the ocean adjacent to Africa, and gradually assume a direction more from the east on longitudes farther west. The continent of Africa does not extend far enough toward the south pole to cause a deflection the contrary way and a gyration of air in the higher latitudes of the South Atlantic, as in those of the North Atlantic.

125. On the east coasts of China and of South Africa, and in some measure on those of New Guinea and Australia, there are similar deflections of the trade-winds by the highlands and mountain ranges of these countries around toward the middle and higher latitudes, where the deflected currents become a part of the easterly currents of these latitudes. For this reason the winds on the southeast of China and the adjacent sea, comprising the Philippine and the Ladrone Islands, are more from easterly and southeasterly directions and are damper than they otherwise would be, and the rainfall is more abundant. This effect, however, is much obscured here by the great disturbing influence of the monsoons, which introduces great

annual changss and alterations in both the directions of the winds and the amount of rainfall. East of New Guinea and Australia the normal direction of the trade-winds is changed to one more from the east and northeast, and on the east coast of South Africa a similar but still greater effect is produced upon the direction of the trade-wind. Consequently the calm-belts and dry zones are very much broken up on all these eastern coasts in the lower latitudes of both hemispheres, and great climatic changes introduced.

The easterly current over the middle latitudes of the North Pacific, strengthened a little by the deflected current on the east coast of China, on arriving at the west coast of North America, is likewise in part deflected by the coast and the mountain ranges, a small part around northward by Alaska on to Behrings Strait, but mostly around to the right along the coast of California and on to the latitudes of the northeast trades, where it becomes a part of the westerly current in the direction of China. This latter deflection gives rise to a predominance of northwesterly and northerly winds along the coast of California and the adjacent sea, and to a more northerly direction of the trade-winds west of Mexico for some distance westward, and to a breaking up of the continuity of the calm-belt and the dry zone which would otherwise exist here on those latitudes.

In the South Pacific Ocean the strong easterly current of the middle latitudes impinges against the high mountain range of the Andes, and a part of it is deflected around by the coasts of Chili and Peru and over the adjacent ocean until it likewise joins the general westerly current in the latitudes of the southeast trade-winds. On this account these winds seem to extend much farther south here and are more southerly than in the trade-wind zone farther west, where the direction becomes more from the east. Here, consequently, the deflected current breaks through the belt of high pressure of these latitudes and interferes with the continuity of the rainless zone of the tropical calm-belt and southeast trade-winds.

126. In consequence of the general circulation of the at-

mosphere, both the temperature and the amount of rainfall are often different on the two sides of a mountain range, especially where the direction of this range is normal to the prevailing direction of the wind. The current, in passing over the mountain, carries the aqueous vapor up to a higher altitude where it is cooled by the expansion of the ascending air and condensed, just as in the case of a vertically ascending current in the equatorial rain-belt; so that on the windward side of the mountain there is an unusual amount of rainfall. This is especially the case if the rain-producing currents pass to the mountain side from the ocean, where the air is generally more nearly saturated, and more especially if it passes from a warmer ocean to a colder district, for then the air has greater capacity for moisture, and is liable to be very nearly saturated before it reaches the mountain side and begins to ascend. On the leeward side the air descends to a lower level, and if it were even saturated with aqueous vapor in crossing the top of the range, it soon becomes unsaturated, and consequently there is no longer any condensation or rainfall, so far as it depends upon the general current passing over the mountain, and there can therefore be no rainfall on this side, unless it arises from other local and temporary causes.

In the general circulation of the atmosphere, we have seen that the average or resultant direction of the winds in the middle and higher latitudes, at the earth's surface and in the lower strata of the air, is nearly from west to east, and in the zones of the trade-winds there is a large west component of motion. Hence in the case of all high mountain ranges extending north and south, or approximately so, especially in the case of coast ranges, the west sides of these ranges, in the middle and higher latitudes, should be the rainy sides, and the east sides the dry sides, and the reverse in equatorial and tropical latitudes. Hence in the northern part of California, in Oregon, Washington Territory, and Alaska, along the Pacific coast, there is an abundant annual rainfall, of from 50 to 75 inches and more, as may be seen from Loomis's rain chart,[11] caused by the coast ranges of mountains, the winds in this case coming directly

from the ocean. After passing the coast ranges in these high and cool latitudes the amount of aqueous vapor left in the air is very much diminished, so that farther on there is but little rainfall, and especially east of the main Rocky Mountain range. There is, consequently, a large area east of this range, extending to about the 100th meridian west from Greenwich, over which the amount of annual rainfall is small. Farther east, over the United States and Canada, the vapor for producing rain, as has been explained, comes mostly from the Gulf of Mexico, and the causes of the ascending currents necessary to give rise to condensation and rainfall depend mostly upon the local and temporary disturbances of ordinary rain-storms, and but little upon mountain ranges.

In western Europe, where the westerly winds from the Atlantic Ocean first strike the continent, especially on the west sides of the mountain ranges, even those near the ocean of only moderate height, and on the west side of the Caucasus Mountains on the east side of the Black Sea, over which the winds pass before reaching them, there is mostly an abundant annual rainfall; while in the interior parts of Asia, especially in Tartary and Mongolia, the amount of annual rainfall is scant, the aqueous vapor having been mostly condensed and having fallen in rain, in passing over the mountain ranges of Europe, and in ascending currents in ordinary rain-storms, before reaching them.

On the west side of the Andes, south of the parallel of about 35°, in Chili and Patagonia, the amount of rainfall is very great, arising from the upward deflection of the moist and strong air currents of these latitudes, coming directly from the ocean; while on the east side of the range of the Andes here, there is scarcely any rain.

27. In the trade-wind latitudes of both hemispheres, where the air, on account of its west component of motion, is deflected upward by the mountain ranges of Central America, and the Andes in South America, the rainfall is very great, while on the opposite sides, especially in Peru, and on the ocean adjacent, there is scarcely any rain, although it is said

that the air here is often nearly or quite saturated with aqueous vapor, and that fogs and mists frequently occur. At the headwaters of the Amazon on the east side of the very high range of the Andes, and where the warm and very moist air currents of the zone of the southern trade-winds strike, the amount of annual rainfall is especially great, but this is increased by the rainfall of the equatorial calm-belt, which oscillates here through a considerable range of latitude, and so increases the annual rainfall over the whole of this range. On the east coasts also of South Africa, of Madagascar, and of Australia, within the zone of the southeast trade-winds, and which on the oceans is a rainless zone, there are considerable rainfalls, while on the west sides there are scant rainfalls, if not an almost entire deficiency, and the latter is the case all the way through the interior of Australia to the western coast.

The large annual rainfall on the eastern coast of China is due in part to the same cause, but mostly perhaps to the summer monsoon which draws the warm and moist air from the adjacent ocean over the mountain ranges in toward the interior of the continent.

On both sides of the Rocky Mountains and the Andes on the parallels from 30° to 40°, there is scarcely any rainfall, for these being the parallels of little or no easterly or westerly motion of the atmosphere, there is no upward deflection of the air on either side, and besides, being in the belt of high pressure, there is a tendency in the air to settle down toward the surface, and hence there is little rainfall.

28. Wherever, on account of the prevailing currents, there is a wet and a dry side to the mountains, the dry side is the warmer. The air in ascending on the wet side, until condensation takes place, cools at the same rate with increase of altitude, as it is warmed on the other side in falling through the same distance. But after condensation commences, as may be seen from Table III, the rate of cooling of ascending air is much less, especially where the temperature is high, as in the lower latitudes; while on the other side, where the air descends from its highest level to that where condensation commences, the rate

is the same as that of dry air, except that it is diminished a little by the evaporation of the cloud-particles immediately after it first begins to descend. Suppose the height of the range is 3000 meters, but that the condensation begins at the altitude of 1000 meters. Then if the rate of cooling on the average, while the air is ascending and condensation is taking place, is 0°.5 C. for each 100 meters of ascent, the air is cooled in this ascent 10°, while in descending on the other side to the same level of incipient condensation in ascending, the increase of temperature is 20°, or nearly, and so there is a gain of nearly 10° C. of temperature from the time the air left the level of incipient condensation on the one side until it descended to the same level on the other side. Since the rate of cooling below this level as the air ascends, is exactly the same as that of increase of temperature in descending on the other side through the same distance, of course there is the same difference of temperature at the earth's surface between the two sides, if the surface on each side has the same level, as there is at the level of incipient condensation as the air ascends. The effect is that of a temporary *foehn* to be considered farther on, and is here a permanent effect continuing through the year, with most probably a considerable annual inequality, since the strength of the general air currents upon which it depends is greater in winter than in summer. Of course the full effect, as computed above, is never reached, since the computation assumes that the air is cooled and heated simply by expansion and compression; but the temperature depends upon a balancing of the rates of absorption and radiation, and the radiation of the warm air on the one side is greater than that of the cold air on the other side, and consequently the difference is a little less than that of the computation. As the air flows away from the base of the mountain range eastward, of course the temperature becomes less until it is reduced to that depending upon the local conditions and independent of the *foehn* effect of the distant mountain.

In accordance with what precedes, there is a belt close along the eastern side of the Rocky Mountain range, with a mean

temperature above that of the same latitude farther east, and which is sensibly felt to a considerable distance over the plain, and this is especially the case in the winter season. Accordingly the coldest winds in winter at Denver and Cheyenne, and other places similarly situated on the east side of the Rocky Mountain range, are not the northerly or northwesterly winds, but those from a northeasterly direction. It is also said that the mean winter temperature at Georgetown, Colorado, close to the base of the mountain, is milder than that of Denver, some forty miles farther east, although the elevation of the former is nearly 4000 feet greater.

A similar effect seems to be produced by the range of the Andes in South America. According to Dr. Gould's isothermal chart of the mean annual temperature of the southern part of South America,[31] the temperature immediately east of the range is several degrees of the Centigrade scale higher than on the west side, and also considerably higher than that farther east on the same parallels, at least north of the parallel of 40°, where the continent is wide, and the isotherms are well determined on the eastern side of the continent.

CHAPTER V.

MONSOONS, AND LAND- AND SEA-BREEZES.

INTRODUCTION.

129. HAVING treated the general circulation of the atmosphere with its annual inequalities or changes with the seasons, arising from the difference of temperature between the equatorial and polar regions and its annual variations, we come now to consider the less general, but still often powerful, atmospheric disturbances arising from differences of atmospheric temperature between continents or islands and the surrounding oceans, or between any region of the earth's surface, abnormally heated or cooled, and the surrounding parts of that surface. In the general circulation of the atmosphere the temperature difference upon which it depends is permanent, though subject to large annual inequalities; but in these secondary and less general disturbances there is little or no permanent tem perature disturbance, independent of the annual and diurnal changes and reverses, and so they are mostly of the latter character, and these give rise to annual and diurnal alternations of direction and velocity of motion. The former of these alternations are called *monsoons*, and the latter *land- and sea-breezes*.

It has been shown, § 68, that in the lower latitudes the mean annual temperature of the continents is greater than that of the oceans on the same latitudes, and the reverse in the higher latitudes. There is, consequently, a permanent difference of temperature between the continents and the oceans in the equatorial regions, the mean temperatures of the former being the greater, but this vanishes in the middle latitudes, and becomes reversed in the polar latitudes. This difference, however, at most, is small in comparison with the great annual changes and

reversals of differences, except near the equator, where these annual changes are small. These permanent differences of temperature must give rise to corresponding permanent atmospheric circulations, with motions of the lower part of the atmosphere from the colder to the warmer region, and the reverse above, to which are superadded the usually greater disturbances depending upon the annual variations of temperature. Monsoons are usually defined to be winds which blow six months in one direction and the other six months in the contrary direction; but this complete reversal takes place only where there is no permanent atmospheric disturbance of any kind, such as that of the general circulation with its various deflections by continents and mountain ranges, or from the permanent part of the monsoon influence just referred to. The monsoons, therefore, are generally not complete reversals of direction with intervening intervals of little or no velocity, but the directions at the different seasons may change less than a quadrant, and the strength of the winds may vary considerably at different seasons, but not entirely vanish. In fact, where there are strong perennial winds, the monsoon influence may be such as to merely change a little the strength and direction of these winds, one way at one season of the year and the contrary way at the opposite season.

130. The temperature differences or gradients upon which the monsoons depend are not those of the absolute temperatures, but the gradients of the differences between these and the normal temperatures of latitude of summer and of winter. From the temperature gradients of the latter or normal temperatures of latitude arise the general atmospheric circulations of the two hemispheres, so that these must not be again taken in as a part of the monsoon influences; but these latter must depend upon differences or temperature gradients in the abnormals of temperature, that is, departures of temperature from the normals of latitude, for without such departures we should have no monsoon influences, but simply the undisturbed general atmospheric circulation, with velocities a little greater in winter than in summer, but no monsoon. For instance, the

interior of a large continent may be much warmer than the polar and much colder than the equatorial side, but still the temperature gradients arising from such differences may not form any part of the monsoon influence. But if there is an increase or decrease from the interior outward of the departures of actually existing temperatures from the normals of latitude, then we have a temperature disturbance which gives rise to a monsoon, or at least to a monsoon influence.

131. Land- and sea-breezes, at least of small islands, are similar to monsoons, the latter being a regular alternation, more or less complete, of the direction of the wind between summer and winter, and the former between day and night; but the lengths of the periods of the alternations are very different. In the case of continents the land- and sea-breezes are sensible along the coasts only, where the temperature gradients are greatest; for the periods of alternation are so short that a complete system of interchanging motions between the ocean and the land, extending into the interior of the continent, cannot be formed and reversed every day, and so the winds in this case generally extend only a short distance either inland or out on the ocean. In the case of a small island, however, where a complete interchanging circulation can be inaugurated in a short time, there is little difference between the land- and sea-breezes and the monsoons of continents, except in their extent and in the periods of alternation.

132. The strength of the monsoon, or of the land- and sea-breeze, depends very much upon the nature of the surface of the continent. In the case of a perfectly flat continent with no highlands or mountain ranges, there would, of course, be an interchange of air between it and the ocean in case of difference of temperature, that of the lower part moving toward the warmer region, and that of the upper part away from it; but the monsoon effects would be comparatively small, and would not at all have the great strength of surface winds which is usually observed. The interchange would be mostly in the great mass of air above, and no strong motion would take place at the earth's surface. In this case, also, the land- and sea-

breezes have but little strength, and are felt near the coasts only, but they are very much increased in strength, and are felt at a much greater distance, where there are hillsides and mountain slopes near the coast.

In the annual and diurnal oscillations of temperature the amplitudes are small on the ocean surface, and in the air at all altitudes above it, and also in the great mass of air over the continent except in a stratum next the earth's surface, of small depth in comparison with that of the whole. The monsoon effects, therefore, depend mostly upon the temperature differences between the continent and the ocean of only a comparatively thin stratum of the atmosphere next the earth's surface, of which the part over the continent is very much heated above, or cooled below, that of the ocean. The temperature differences of such a stratum, over a perfectly level continent, even if they were very great, would give rise to very little horizontal disturbance of the atmosphere. If this stratum over the continent were greatly heated, it might give rise to the unstable state, from which would result numerous, but very small, local eruptions through the strata above, but no sensible monsoon effects. On the other hand, if it were cooled down to a very low temperature, the increased density would tend mostly to keep it next the earth's surface, and there would be scarcely any tendency to flow away laterally toward the warmer ocean. But if the surface of the continent is convex, or if it has highlands with long slopes, or the interior is in almost any way considerably elevated above sea-level, the tendency in the case of the summer monsoon to flow in from the ocean toward the interior of the continent, or the reverse in that of the winter monsoon, is very much increased. The same is true with regard to land- and sea-breezes where there are mountain elevations near the coast.

133. The reason of this can, perhaps, be made clearer by assuming an analogous case, but one which will be more easily understood. If a perfectly level plain were covered with a stratum of water, there would, of course, be some flow of water in all directions from the interior outward; but unless the depth

were considerable in comparison with the extent of the plain, the current might not be perceptible except near the outer part. But if the same depth of water were placed upon the surface of a continent with an elevated interior having long declining slopes from the interior outward, the rapidity of the rush of water would be very great, and somewhat in proportion to the steepness of these slopes. The case is somewhat similar where there is a stratum of air next the earth's surface abnormally cold and dense in comparison with the air above it and over the ocean. If the earth's surface is flat there is little tendency in this air to move in any direction, but if it rests upon declining slopes, this tendency, as in the case of water, is very much increased. Of course the absolute tendencies in the case of water and the abnormally dense stratum of air are very different, the former being due to the force of the whole pressure of the fluid down an inclined plane, but the other merely to the increased pressure arising from the diminished temperature and increase of density. The relative tendencies, however, for different surfaces are the same in both cases, and somewhat in proportion to the steepness of the declining surface.

The very reverse of this is the case where the stratum of air next the earth's surface is abnormally heated in comparison with the temperature of the air above and around about. If the excess of temperature in this case is the same as the deficiency in the other, the tendency now to flow from all sides *toward* the interior *up* the slopes is exactly equal to the tendency to flow *outward* and *down* these slopes in the other case. There being now a deficiency of pressure in the warm rarefied stratum, where this rests upon an inclined surface, the greater pressure of the heavier air everywhere else at the same level tends to drive the lighter surface air up the slopes with the same force with which it descends in the other case, both depending upon the same difference of density between the stratum and the air generally at the same levels.

The case of the warm and rarefied air ascending on the slopes is somewhat similar to that of warm air ascending in a flue. If

the flue is horizontal, although the air within is kept much warmer than that of the flue, and the outside air generally, the warmer air gradually, but very slowly, passes out of the flue and allows other air to take its place; but if it is only a little inclined, the air ascends the incline and escapes rapidly from the higher end, and the more so the greater the differences of temperature, the length of the tube, and the steepness of the inclination. So the tendency of the warmer stratum of air to flow up an inclined surface increases with increasing differences of temperature, and increasing length and steepness of slope.

134. There is also another consideration in connection with the subject of the monsoon influence of highlands. The tendency of air to ascend or descend and to give rise to ascending or descending currents depends upon differences of temperature between the air and that of the surrounding regions at the same level. But it is a matter of observation that the temperature of highlands, and especially of high plateaus, in summer, is nearly as great as that on plains near sea-level. The temperature, also, for the altitudes above the surface, at least to a considerable height, must be much greater than that of the surrounding air at the same levels, since the rate of decrease of temperature with increase of altitude above the surface of the plateau is somewhat the same as above any plain near sea-level. If a portion of air, therefore, either on a horizontal plane or a slope near sea-level is only a little warmer than the surrounding air on the same level, it tends to ascend and to give rise to an ascending current; but if air at the same temperature is high up on some mountain side or plateau, this tendency is much increased, because now the difference of temperature between this air and the surrounding air at a distance on the same level is much greater.

If a tall flue with a temperature only a little raised above the surrounding temperatures at sea-level were elevated to the top of a high mountain, where the surrounding air is much colder and more dense, the draught of the flue would be very much increased. So the wide column of air of higher temperature over a high plateau and extending up to a considerable height above

the surface has a much greater tendency to ascend than a similar one of the same temperature on a low plane near sea-level.

This effect, however, is felt mostly in the summer monsoon influence, for in winter the temperature of the plateau is not so much below surrounding temperatures as it is above them in the summer.

135. In accordance with the preceding view of the principal cause of monsoons and land- and sea-breezes, it is seen from observation that all the great monsoons and the strongest land- and sea-breezes are found—the former in countries and on oceans adjacent to high mountain ranges, and the latter along coasts with high mountains in the background. Neither the heated interior in summer of the Great Sahara of northern Africa nor of Arabia and Persia, which is considered the warmest region on the globe, causes, during this season of the year, any considerable indraught of air. It is true that at this season the northwesterly winds prevail a little more on the northwest coast of Africa and the ocean adjacent, due, no doubt, to the influence of the highly heated desert of the Sahara; but over Arabia and Persia the northwest winds continue to blow almost incessantly during June and July, away from the heated interior toward the Arabian Sea, though at times they are quite light. Even at this season the tendency to flow in towards this heated district is not sufficient to overcome the northwest wind of this region arising from the deflection around the elevated regions of Asia (§ 123). The monsoon influence, therefore, of countries mostly level without an elevated interior, however highly they may become heated in summer or cooled in winter, is not very great.

With regard to land- and sea-breezes Mr. Laughton says:[32]

"This seems to be sufficiently established by facts, that the sea-breeze nowhere blows with any great strength, except where there are mountains in the background; and that these mountains are, during the afternoon, when the sea-breeze has gotten well home to them, constantly enveloped in mist or storm. The Blue Mountains of Jamaica afford the best illustration of this; and it is on the coast of Jamaica, more especially of Port Royal, that the sea-breeze has a force unknown anywhere else."

In the same connection it is said:

"As a matter of geographical fact, the land-breeze has nowhere any noticeable strength unless there are mountains in the immediate neighborhood."

136. In the case of the summer monsoon, where the interior of the country is so elevated that the current ascends the slopes to an altitude where condensation of the vapor takes place, the latent heat of condensation adds much strength to the current, just as in the case of the trade-winds, in which their strength is increased by the latent heat of condensation in the equatorial rain-belt. On this account the summer monsoon of the North Indian Ocean is much stronger than the winter monsoon: so much so, that the southwest monsoon is often spoken of as *The Monsoon*, the northeast monsoon being insignificant in comparison with it. Notwithstanding the trade-wind is combined with the monsoon effect, the resultant of both produces in the Arabian Sea only a gentle and steady breeze during the winter season; whereas the southwest monsoon of the summer is a steady gale of so great strength that it is impossible for even steamers to force a passage from Bombay to the Gulf of Aden in June and July.

In all cases, also, of extraordinarily strong sea-breezes, there are high mountain elevations near the coast on which there is a vast amount of condensation and precipitation of rain at the time.

THE MONSOONS OF ASIA AND THE INDIAN OCEAN.

137. In no other part of the world are the conditions for a powerful monsoon influence so favorable, and the strength of the monsoons produced so great, as in India and the North Indian Ocean. The Himalayas on the north with an east and west extent of 1300 miles and an average height of 18,000 feet, the parallel Kuenlun range still farther north, much greater in extent but of less height, with the high plateaus of Thibet and Cashmere between, and beyond these yet the vast and elevated deserts of the interior of Asia between the Kuenlun and Altai

ranges, together with the warmth and great capacity of the air for moisture over India and the North Indian Ocean, for reasons just given in the preceding pages, all combine to add strength to the monsoons, especially those of the summer season. As the sun in early summer approaches the northern tropic, the air of the Himalayan southern slopes and the desert of Gobi and the sunburned plains of Central Asia becomes warmed up to a temperature much above that of the adjacent and surrounding air at a distance on the same level, and so a powerful centripetal and ascending tendency is induced, by which the air is drawn in toward the centre of warmth and rarefaction from all sides, but especially from the equatorial side, where the influence is felt even beyond the equator. For there is not only the tendency of the heated stratum of air in contact with the southern slopes to ascend, but the air over the whole of the elevated part of Central Asia, not only near the surface, but up to a considerable altitude, being warmer than that of the surrounding regions on the same level, it has a strong tendency to rise up and leave a partial vacuum into which air from all sides flows to take the place of that which is continually ascending.

The warmth of the air, and consequently the monsoon influence, is much increased in summer by the latent heat given out in the condensation of the aqueous vapor in the almost or quite saturated air at the beginning of its ascent, for, being of high temperature over India and the adjacent ocean, its capacity for moisture is very great, and this is condensed, not only as it ascends the southern slopes to the general height of the Himalayas, but still beyond in its further ascent over the plateaus of Thibet and Cashmere, where the amount of rain during the summer monsoon is often enormous. Although there would, no doubt, be a considerable monsoon influence in the case of perfectly dry air, yet the principal part of the energy in the summer monsoons of India is probably in the aqueous vapor of the air. Some idea of the difference of temperature between the ascending air up the slopes of the Himalayas and that over Thibet beyond, and the air of the adjacent and surrounding regions at the same level, may be

formed from an examination of Table III. From this it is seen that the average rate of decrease of temperature with increase of elevation in ascending air, even up to great altitudes, would be only about 0°.4 C. in each 100 meters, while we know that the rate of decrease in air generally in the summer season, especially in the lower part of it, is much greater, and consequently the difference of temperature above, between the ascending air and the surrounding air at the same level, must be considerable.

188. During the month of April the weather is unsettled, and the wind variable, without any prevailing direction. By the middle of May the summer monsoon, with its accompanying rains, is fairly set in, and soon after arrives nearly at its full strength. The general direction of the wind on the Arabian Sea is from the southwest, or rather from the west-southwest, and is here called the *Southwest Monsoon*. The direction is more from the west than it otherwise would be, both on account of the deflected current around the highlands of Asia, § 123, and also on account of the deflecting force of the earth's rotation, which, however, is not very strong here, so near the equator. Further east, in the Bay of Bengal, the wind is more southerly; along the coast of China, and over the Chinese Sea, it is from the southeast, and in Japan it is still more easterly. Along the whole of the north coast of Siberia the wind blows inland during the summer season, though not directly in towards the interior, but rather from the northwest, the direction being the resultant of the monsoon influence and the general easterly tendency of the atmosphere in these latitudes. The prevailing direction of the wind observed in the Vega expedition in summer from North Cape to Yokohama was from the northwest. The monsoon influence then is felt on the north side of Asia; but, as we would expect, it is comparatively weak, because there are not many highlands and high mountains in Siberia, and the effect arising from the condensation of aqueous vapor is small.

The general tendency in the summer season is to flow in from all sides toward the heated interior, with a slight deflec-

tion to the right, on account of the influence of the earth's rotation, and consequently to flow out above in all directions. In consequence, however, of the high range of the Himalayas, little air from India passes over this range toward the north; but it continues mostly to ascend after arriving at the top of the slope, until it is immerged into the returning current above toward the equator. It is said, however, that there is some air passing through all the highest passes of the range, on toward the interior of the continent. According to the experience of Prjivalsky in his travels over the high plateau of Thibet in June and July, there must be a great deal of air brought into this region at this season, which in ascending gives rise to much rain. It is said: "As to rain, it poured down every day,—sometimes several days without interruption. The amount of vapor brought by the southwest monsoon, and deposited there, is so great that, during the summer, northern Thibet becomes an immense marsh." But this vapor comes in, most probably, from other directions, and not from the southwest monsoon only, since southern Thibet, next the Himalayan range, is said to be comparatively dry.

In the Arabian Sea the monsoon influence in summer is in the right direction and sufficiently strong to entirely overcome and reverse the northeast trade-wind; but in the Chinese Sea, and the vicinity of the Philippine Islands, the effect is such as to change the northeast trade-wind to a southeast wind. Not only is the regular trade-wind systems, then, of the North Indian Ocean entirely broken up in the summer season, but the monsoon influence draws air from the other side of the equator, thus strengthening the southeast trade-winds there. This is especially the case over all the equatorial region between China and Australia, since, while the interior of Asia during its summer is heated and the air rarefied, that of Australia is cooled and its pressure-increased, so that the temperature and pressure gradients between the two countries are due to both causes. There is consequently no equatorial calm-belt during the monsoon; but the air is drawn from about the parallel of 25° S. across the equator to the rarefied central part

of Asia—not directly, but is deflected toward the west, south of the equator, and then eastward, after crossing the equator to the north side, by the deflecting force of the earth's rotation. The region of greatest temperature, or, at least, of greatest departure from the normals of latitude, upon which the monsoon depends, and lowest barometric pressure, is not now near the equator, but in the interior of Asia, and the highest pressure is in the southern belt of high pressure, about the parallel of 30° S., so that there is now a continuous gradient of pressure decreasing from this belt across the equator into the interior of Asia. There is therefore, at this season of the year, an enormous disturbing influence, due to this monsoon, which entirely breaks up the general circulation of the atmosphere and the tropical and equatorial calm-belts.

139. As in the case of the general motions of the atmosphere, so in that of monsoons—wherever the vapor-laden current strikes a range of hills or mountains, and is deflected upward, there is a very large amount of rainfall during the time. The current of the southwest monsoon, in its passage from the Arabian Sea toward the southern slopes of the Himalayas, first encounters the western Ghauts, a range of mountains from 4000 to 6000 feet in height, near the Malabar coast, where a great part of its vapor is condensed, and the rainfall is immense. Passing on then with little precipitation to the mighty Himalayas, it now in the ascent loses nearly all of the remaining part of its vapor, which, falling on a small area on account of the steepness of the slope, causes an immense amount of rain to be precipitated, greater or less according to the nature of the country and the height and steepness of the mountain sides.

But the greatest rainfalls are experienced in the mountains of Khasia, north of the head of the Bay of Bengal, where the warm and moist air of the nearly equatorial winds blowing across the bay is somewhat concentrated by the converging coasts and highlands on the east and west before it commences its ascent up the mountain slopes, and having previously lost little or no vapor by condensation, as it does in the case of the

southwest monsoon in passing over the western Ghauts, the whole vapor is now condensed in the ascent of the Himalayan slope. According to Dr. Hooker, as cited by Laughton:[18]

"The climate of Khasia is remarkable for its excessive rainfall. Attention was first drawn to it by Mr. Yule, who stated that in the month of August, 1841, 264 inches fell, or twenty-two feet; and that during five successive days, thirty inches fell in every twenty-four hours. Dr. Thomson and I also recorded thirty inches in one day and night; and during the seven months of our stay upward of 500 inches fell; so that the total annual rainfall perhaps greatly exceeded 600 inches, or 50 feet, which has been registered in succeeding years. From April, 1849, to April, 1850, 502 inches fell. This unparalleled amount of rainfall is attributable to the abruptness of the mountains which face the Bay of Bengal, from which they are separated by 200 miles of Jheels and Sunderbunds."

On account of the vast amount of precipitation during the summer monsoon over India, and especially the slopes of the Himalayas, the southwesterly monsoon of the Arabian Sea and western India and the more southerly monsoon farther east, is often called the *Wet Monsoon.*

140. The air which is drawn across the equator during the summer monsoon from the southern hemisphere of course simply strengthens the regular southeast trade-wind which exists when there is no disturbance from the monsoon influence; but at and a little north of the equator the wind becomes southerly, and still farther north of the equator it is deflected around by the influence of the earth's rotation into a northeasterly direction, and becomes the southwest monsoon. This is clearly shown by the following table given by Woeikoff,[33] of percentages of the winds from the different directions as deduced from the observations on the ships of Holland, during the months of June, July, and August:

Latitude.	E. Long.	N.	N.E.	E.	S.E.	S.	S.W.	W.	N.W.	Calms.
10°—15° S.	80°—90°	1	3	17	63	12	1	1	0	2
5—10	80—90	3	9	22	40	10	8	7	2	2
0—5	75—85	1	3	22	23	15	12	14	5	5
0—5° N.	80—90	1	0	0	5	18	43	24	7	2
5—10	80—90	0	1	1	3	9	57	23	2	3
10—15	85—95	0	0	0	2	10	64	13	3	9

From this table it is seen that the southeast trade-wind gradually changes around into the southwest monsoon, the observed directions being mostly southeasterly south of the equator and southwesterly north of it, and there is apparently no calm-belt at or near the equator.

141. Toward the end of October, or the beginning of November, the interior of Asia and the slopes of the Himalayan range are cooled down to the mean temperatures of the year, and there has been an equalization of the air-pressure, and the monsoon has subsided; or rather, the pressures have changed to the normal pressures for the mean temperature of the year, and the winds to those of the corresponding general circulation, with high pressure and calms near the two tropics, and low pressure and calms near the equator. But now a reversion of the summer conditions gradually takes place. During the winter season of the northern hemisphere the interior of Asia, and especially of Siberia, becomes cooled down by the radiation of heat into space, through the now very dry air, to a very low temperature, far below that of the normals of latitude for the season, and the density of the air and the barometric pressure are very much increased, instead of being diminished as in the summer season. The effect is now a complete reversal of all the motions between the central and surrounding regions, and instead of a flowing in below from all sides toward the central part, and a flowing out above in all directions, the flow is outward below and inward above. The southern slope of the Himalayas is now cooled by radiation far below the general temperature of the air at the same levels at a distance from it, and the stratum of air in contact becomes dense and heavy, and consequently tends to flow down toward the lower levels of India and the Indian Ocean. The monsoon is now completely reversed, and instead of S.W., S., and S.E. winds around the southern and eastern coasts of Asia, there are N.E., N., and N.W. winds.

The wind over the Arabian Sea and India depending upon the monsoon influence being mostly a northeast wind combines with the regular trade-winds of these latitudes, and the result-

ant is a stronger wind than that which would arise from the monsoon influence alone, but still not nearly so strong as that of the southwest monsoon. It is called the *Northeast Monsoon*, and the monsoons generally of the winter season of the region south and east of Asia are often called the *Winter Monsoons*. Since the air of these monsoons comes from the elevated dry and cool interior of the continent into warmer regions of lower level, it is very dry, especially that in India which comes down from the Himalayas. Hence these monsoons are often called the *Dry Monsoons*.

But when the northeasterly monsoon does not come directly from the elevated and dry interior of the continent, but blows a long distance over the sea, on coming in contact with mountain ranges, as in Ceylon, it gives rise to just as much rain, or nearly so, as the southwest monsoon. The latter is usually a little the stronger, and blowing from more southern latitudes may have a little more vapor, and so for both reasons give rise to a little more rain.

In the summer or wet monsoons we have the effect of the latent heat of condensation in connection with that of the heating from the sun's radiation, but in the winter and dry monsoons we have simply the effect from greater cooling by radiation, which is about equivalent only to that from the greater heating in summer from solar radiation, so that the effect is not nearly so great in the case of the winter monsoons as in that of the summer monsoons in which much energy arises from the latent heat of condensation of aqueous vapor. Accordingly the northeast monsoon, though strengthened by the regular northeast trade-wind, is not nearly so strong as the southwest monsoon, as has been already stated in § 136.

During the winter season the winds of Northern Siberia, especially near the coast, are mostly southerly, blowing from the colder continent toward the warmer polar sea, and hence in a direction nearly the reverse of that of the polar winds of the summer season. There is, therefore, a monsoon along this coast with alternating winds of no great strength; but these are not exactly opposite in direction, but rather from the

northwest in summer and the southwest in winter, owing to the general prevalence of westerly winds in these latitudes arising from the general circulation of the atmosphere.

"In winter the whole coast of Norway has monsoon winds, blowing from the land to the sea: they are N. and N.E. at Christiania; S.E. at Christiansund, Bosekop, and Hammerfest; and S.W. at Vardöe. In summer the conditions are reversed. In northern Norway the winds are variable in summer, and decidedly from the S. in winter." [20]

142. On account of the great lowering of temperature in Asia in winter below that of the normals of latitude, which, considered alone, gives rise in the general circulation to a belt of high pressure about the parallel of 35°, and one of low pressure at or near the equator, this regular system is now completely broken up, and the highest pressure is found in the interior of Asia, and there is a gradient of decreasing pressure thence to about the parallel of 12° south of the equator in the Indian Ocean. The normal equatorial belt of low pressure and calms is now transferred to this parallel, where the northerly monsoon of Asia, crossing the equator, meets the regular southeast trade-winds, and hence the belt of these trade-winds is now confined within the limits of the parallels of 12° and about 25° or 30° S. After the northeasterly and northerly winds of the winter monsoon pass over the equator they are deflected to the east by the force depending upon the earth's rotation, which is towards the left in this hemisphere, so that there is a belt between the equator and the parallel of 12° S., where the prevailing direction of the wind is from the northwest. This is called the *Northwest Monsoon.* Where this monsoon meets the southeast trade-winds we have the conditions of the equatorial calm-belt, a meeting of counter currents, calms, an ascending current over a belt of several degrees in width, and consequently an abundance of condensation and of rainfall over this belt. Espy says: [34] "The monsoon in the Indian Ocean, which blows from the northwest during the summer, terminates in a belt of rains 6000 miles long."

On the western side of the southern Indian Ocean the northwest monsoon is much modified by the whirl of atmosphere

deflected by the high table-lands and mountain ranges of the east coast of Africa, § 125, by which the direction of the wind is changed and the monsoon is changed to what is called in this region the *Northern Monsoon.* According to Espy, "the northern monsoon, which blows into the northern end of the Mozambique Channel in the southern summer, terminates there in a great rain, and the southern trade, which blows into the same channel at the same time, terminates there in the same rain; and the ship that is caught in that channel at that season finds the wind dead against her at whichever end of it she tries to escape." This rainy portion of the channel at this time is simply the western end of the general rainy belt of the Indian Ocean explained above and referred to by Espy, but moved here a little farther south than it is farther east, by the atmospheric whirl referred to above.

143. There are monsoon influences arising from the annual alternations of temperature in Australia similar to those of Asia, but on account of its smaller extent and the absence of very high plateaus and mountain ranges, these influences are comparatively small. In Asia the monsoon influence, at least at the extremes, is the predominating one, giving rise to atmospheric disturbances and velocities of air currents by far greater than those of the general circulation of the atmosphere, so that the latter merely cause perturbations in the regular annual alternations and inversions of the monsoon. In the case of Australia, however, the monsoon influence is the weaker, and its effects merely cause perturbations in the general circulation of the atmosphere. The continent lying mostly within the belt of the southeast trade-winds, these in its vicinity are considerably modified all around the coast of the northern half of it, and also the prevailing westerly winds in the region on, and adjacent to, the southern coast.

During the summer of the southern hemisphere, when the earth is in perihelion, the whole land, exposed to an almost vertical sun at midday, and lying mostly in the southern zone of high pressure and dry air (§ 120), becomes intensely heated, far above the temperature of the surrounding ocean. Conse-

quently the tendency now is for the air below to flow in from all sides toward the central part, and of course the contrary above. In this flow, however, the currents are deflected a little to the left by the force depending upon the earth's rotation, but this force is small on the northern side on account of its nearness to the equator. On the northeast side of Australia, where the southeast trade-winds prevail, their directions are now changed, so that they become easterly and northeasterly winds, especially during the day. On the eastern side along the coast of New South Wales, which is in the south tropical calm-belt, where calms prevail when there is no monsoon or other abnormal atmospheric disturbance, the wind blows from the ocean toward the interior of the continent. Along the southern and southwestern coast the westerly and northwesterly winds which prevail in these latitudes are a little changed in direction by the monsoon influence and blow from a more southerly direction. On the northwest coast, in the zone of the regular southeast trade-winds, the monsoon influence is not strong enough to reverse them, even in midsummer, so that here the wind at all seasons blows from the land toward the ocean, but with a strength a little diminished by the monsoon influence. The effect of this is to bring the line of meeting between the northwest monsoon and the southeast trade-winds a little farther south than it is over the South Indian Ocean generally, where this monsoon influence is not felt, so that in the ocean adjacent to the north and northwest coast this line is brought down several degrees south of the parallel of 12°, and passes over the northern extremity of the continent.

144. The climatic effect of the summer monsoon influence in Australia, where it is not counteracted by the regular southeast trade-winds, is to bring in cool and moist air from the cooler ocean into the dry and heated interior, and so to modify the intense heat along the coast and to increase the rainfall. Accordingly on all sides from the north around by east to the southwest side, where the wind blows from the ocean inland, there is an abundance of rainfall, as may be seen from Loomis's chart.[11] But for this influence, along the coast of New South

Wales, where otherwise calms would prevail, there would be the same scant rainfall, or nearly so, as in the interior on the same latitudes. This rainfall all along the eastern side is much increased, though reduced to a much narrower strip, by the succession of mountain ranges from 3000 to 5000 feet high near the coast.

The abundance of rain on the northern extremity of Australia in summer is due to the line of the meeting of the southeast trade-winds with the northwest monsoon, and the consequent accompanying rain-belt, passing over it at this season; for it is simply the normal equatorial rain-belt brought down to this parallel, mostly by the Asiatic monsoon influence, but in part also by that of Austràlia, as has been explained above.

Along the west and northwest coast of Australia there is little rain, because here the regular southeast trade-winds prevail at all seasons, bearing away the warm and dry air of the interior toward the ocean; for if there were even mountain ranges here to cause upward deflections, the amount of moisture in the air coming from the interior of the continent would be too scant to give rise to much rain.

145. During the *winter* of the southern hemisphere the conditions are mostly reversed, and there is now a tendency of the air to flow *out* below in all directions from the interior toward the ocean. The effect of this along the northeast coast, where the southeasterly trade-winds blow, is to change them somewhat to more southerly winds, which are found to prevail here at this season. Further south, however, along the coast of New South Wales, in the belt of calms and high pressure, and where easterly winds prevail in summer, there are now westerly winds, so that here there is a somewhat regular monsoon, but of no great strength. The monsoon effect upon the prevailing westerly winds along the south and southwest coast is to cause them to blow from a direction a little more from the northwest. Along the whole of the northwest coast and over the whole northern half of the continent there are, at this season, the regular southeast trade-winds, but strengthened somewhat by the monsoon influence.

The climatic effect of the winter monsoon influence is to increase the dryness and to diminish the amount of rainfall over the whole of the interior, which, aside from this influence, is a very dry region, from its lying mostly in the south tropical zone of high pressure and in the zone of the trade-winds, in both of which there is, in general, little rainfall.

Since the mean temperature of land in tropical and equatorial latitudes, for reasons given in § 68, is greater than that of the ocean on the same latitudes, the strength of the winter monsoon influence is less than that of the summer. For if we suppose that, in the annual oscillation of temperature, this rises in summer as far above the mean annual temperature as it falls below in winter, then the difference between the land temperature and that of the ocean must be greater in summer than in winter, since the mean temperature of the land is higher than that of the ocean, and hence the air-pressure is decreased more below the annual mean of the ocean during the Australian summer than it is raised above it in winter.

THE MONSOONS OF AFRICA.

146. For reasons given in the case of Australia, the mean annual temperature of the whole of equatorial and tropical Africa is higher than that of the oceans on the same latitudes. Since the monsoon influence depends upon the range of the annual oscillation of temperature, and this is not very great here, so near the equator, this influence is not so great here as it otherwise would be: in fact, there is little of the real monsoon influence, which would cause a reversion of directions at the two seasons; for the mean temperature of the land being considerably greater than that of the ocean, it scarcely becomes lower in general than that of the ocean in winter. The consequence is, that while there is a strong tendency in the air at the earth's surface to flow in from the ocean in summer, or at the times of highest temperature, there is little or no tendency to flow out when the temperature of the land is reduced to the lowest, since it is then but little below the temperature of the adjacent

ocean. Over the whole of the northern part of Africa, and even over the Mediterranean Sea, the regular northeast trade-wind prevails, with slight annual variations and strengthened a little by the summer monsoon influence, throughout the whole year; but along the northwest coast of Africa, and to a considerable distance over the adjacent part of the Atlantic, it becomes northerly and even northwesterly in midsummer, the air being drawn in towards the heated region of the Sahara. At the same time in the Gulf of Guinea and to some distance into the Atlantic, the regular southeast trade-wind is changed, first to a southerly, and nearer the coast to a southwesterly wind, called the *Southwest Monsoon of Africa*, which is really not a monsoon, but simply a change in direction of the regular trade-winds by the indraught of air toward the Sahara at this season. But at the opposite season there is little reversing action, since there is then but little difference between the temperature of the land and that of the ocean, and so the southeast trade-wind then has nearly its usual direction.

147. West of the Gulf of Guinea and the coast of Liberia, and a little north of the equator, there is a wedge-shaped area with its apex extending far west into the Atlantic Ocean, in which calms and light variable winds usually prevail, and the winds, especially in the summer season of the northern hemisphere, sometimes blow directly in toward the interior of Africa. This is caused by the nearly permanent difference of temperature between the equatorial parts of the continent and the ocean; for the annual variations here are small, and consequently there is very little real monsoon effect, or reversal of the winds with the seasons. The tendency is to draw the air in from the ocean toward the Sahara, and this entirely counteracts the regular northeasterly trade-winds in summer up to the parallel of about 13°, causing an apparent widening of the equatorial calm-belt in this region. During the winter of the northern hemisphere, when the Sahara becomes cooled down somewhat and the region of greatest heat and rarefaction of the air is transferred to South Africa, the northeasterly trade-winds blow again to a lower latitude, as in other parts, and there is

now a considerable narrowing of this area. The tendency now is to draw the air more toward South Africa, so that this area not only becomes narrower, but there is now often a northwesterly wind blowing from this region into the Gulf of Guinea.

There is a wide zone through Central Africa, comprising the country of the Soudan, and lying mostly between the parallels of 5° and 15° N. and extending eastward to the Abyssinian mountains, in which there is apparently a somewhat regular monsoon, the winds changing with the seasons from a northeasterly or northerly direction in the winter to a southerly direction in the summer. During the winter the regular trade-winds, which are always more northerly on land, where there is more friction, than on the ocean, where there is less, blow down to the parallel of about 5° N. But during the summer the southeasterly trade-winds, becoming southerly winds at and near the equator on account of there being here no deflecting force from the earth's rotation, blow up to about the parallel of 15° N., and hence between these parallels there is an annual reversion of the directions of the winds from northerly winds, or nearly so, in the winter, to southerly winds in the summer, of the northern hemisphere, and back again. This, however, arises mostly from the general oscillation of the equatorial calm-belt all around the globe, depending upon varying temperature conditions extending over the whole of each hemisphere, as has been explained, but also in some measure upon the annual transfer of the local more heated regions of Africa from the northern to the southern hemisphere and back again. The same is observed more or less in nearly all parts of the zone over which the equatorial calm and rain-belt oscillates.

MONSOONS OF NORTH AMERICA.

148. On the continent of North America we have monsoon influences similar to those of Asia, but not nearly so strong, because the extent of the continent, and consequently the annual range of temperature, are not so great. They are, for the most part, not sufficiently strong to completely overcome and

reverse the current of the general circulation of the atmosphere, and so to produce a real monsoon, but they cause great differences between the prevailing directions of the winter and summer winds.

In the summer the whole interior of the continent becomes heated up to a temperature much above that of the oceans on the same latitudes on each side—indeed above that of the Gulf of Mexico and the Pacific Ocean on its southern and southwestern borders. The consequence is that the air over the interior of the continent becomes more rare than over the oceans, rises up, and flows out in all directions above while the barometric pressure is diminished, and the air from all sides, from the Atlantic on the east, the Pacific Ocean on the west, the Gulf of Mexico on the south, and the polar sea on the north, flows in below to supply its place. On the east the tendency to flow in is not strong enough to counteract the general easterly motion of the air at the earth's surface in the middle latitudes, and to cause a westerly current, but it simply retards the general easterly current and gives rise to a greater prevalence of easterly winds along the Atlantic sea-coast during the summer season. On the southern and southeastern coast, in connection with the deflection referred to in § 123, it causes the prevailing winds to be southerly and southeasterly instead of northeasterly, as they otherwise would be in these trade-wind latitudes. It is precisely the same effect as is produced in the region southeast of China (§ 138). The monsoon influence in the Mississippi valley and westward is much strengthened by the gradual slope from this valley up to the high plateaus east of the Rocky Mountain range, for reasons which have been already given (§ 132); so that when this slope in summer becomes heated the surface air tends to flow up it toward the mountain range, and causes winds, which otherwise would be southerly ones, to become more southeasterly, and southwesterly winds to become southerly ones.

In winter the thermal conditions over the continent are reversed. The interior of the continent is now the coldest part, and it is especially colder than the surrounding oceans at

that season. It has also very high plateaus and mountain ranges. The air, therefore, of the lower strata, and especially those next the earth's surface, now tends to flow out in all directions to the warmer oceans and the Gulf of Mexico, and especially to run down the long slope of plateau from the Rocky Mountains into the Mississippi valley. The effect over the whole of the United States east of the Rocky Mountains is to cause the winds, which otherwise would be westerly and southwesterly, to become generally northwesterly winds, instead of southerly and southwesterly ones, as in summer. There is not a complete monsoon effect, but simply a great change between summer and winter in the prevailing directions of the winds. In Texas, however, and farther east along the northern border of the Gulf, the effect is somewhat that of a complete monsoon. In New England and farther south in the Eastern States the monsoon effect is to cause the prevailing winds to be from some point north of west, instead of south of west as in summer.

149. In summer Central America and Mexico have a much higher temperature than that of the adjacent tropical sea on the southwest, and having high mountain ranges and elevated plateaus, there is consequently a strong tendency to draw in air from the southwest at this season, which not only entirely counteracts the regular trade-winds of these latitudes, but even reverses them and causes southwest winds. The effect is to cause in midsummer a large area here, extending far westward, of calms and irregular and light winds, mostly southwesterly ones, and an apparent widening of the equatorial calm-belt at this season so as to make its northern limit reach up, along the coast, nearly to the parallel of 20°. The effect is similar to that in the Atlantic west of the Gulf of Guinea and Liberia, except that it here appears to be some greater, and causes a true monsoon effect, since during the winter the regular northeasterly trade-winds prevail, but strengthened by the reverse thermal conditions of the winter season. On the eastern side and over the western end of the Gulf of Mexico there is a somewhat regular monsoon effect, the prevailing winds being

easterly, or blowing toward the land, during the summer, and the reverse in winter.

Along the west coast of North America in the middle latitudes there is a strong monsoon influence; for the interior of the continent becomes heated in summer to a much higher temperature than that of the southwesterly ocean, and hence a strong current is drawn in from this direction, at right angles to the general trend of the coast, which, combining with the general southwesterly winds of these latitudes in the general circulation of the atmosphere, causes the strong and steady westerly and southwesterly winds of this region during the summer. Farther north, up toward Alaska, the summer monsoon effect is combined with the current caused by the deflection of the continent (§ 123) as well as the general easterly current of high latitudes, so that the winds here are generally southerly, but still have somewhat of a monsoon character, being southerly and southwesterly in summer, and easterly and southeasterly during the winter.

All along the northern coast of America, as along that of Siberia, the monsoon tendency is to draw the air from the colder land to the warmer ocean in winter, and the reverse in summer; and these effects, combined with the general easterly motion of the atmosphere in these latitudes, gives rise to prevailing southwesterly winds in winter and northwesterly ones in summer. The winter monsoon influence, however, is small here—much more so than in Siberia, for the ocean contains so many large islands that it has rather a continental than an oceanic winter temperature; and besides, it has not the influence of a warm current—such as the continuation of a part of the Gulf Stream along the northern coast of Europe and Asia.

MONSOONS OF SOUTH AMERICA.

150. In South America we have conditions similar to those of Australia, and the monsoon influences are well marked, being stronger than those of Australia, both because the continent is larger and the mountain ranges higher. Along the northeast-

ern coast of South America, and in the whole Amazon valley, the indraught of air from the ocean toward the interior of the continent and the eastern slope of the Andes is so strong during the summer of the southern hemisphere, that there is now a continuous northeasterly wind from the region of the northeasterly trade-winds, far into the interior, and the monsoon effect appears as a continuation of these trade-winds across the equator into the opposite hemisphere. But this is somewhat of a perennial effect, though much greater during the summer of the southern hemisphere than at the opposite season. For the mean temperature of equatorial and tropical South America being considerably greater than that of the oceans, and the annual thermal changes being small, there is perhaps scarcely any time when the land is colder than the ocean, and when there is a reverse monsoon effect.

The effect of the monsoon influence in summer over Brazil, which is mostly in the zone of the southeasterly trade-winds, together with that of the deflection by the continent, and especially the range of the Andes, is to change these winds into northeasterly and northerly ones, which are the prevailing winds here in the summer. The effect is similar to that in the southern part of the United States and the Gulf of Mexico, by which the winds here are changed from the regular northeasterly trade-winds to southeasterly and southerly winds in the summer season. During the winter the winds are more variable here, since the monsoon influence rather antagonizes the easterly and northerly winds which arise from the deflection of the continent and the Andes, but they are largely polar and southwesterly winds, as in the corresponding part of the United States they are polar and northwesterly winds.

On the coast of Chili and the adjacent part of the ocean the prevailing winds are mostly southerly and southwesterly in December, January, and February, the southern summer, but northerly and northwesterly at the opposite season of the year. This arises in part from the deflection of the strong westerly winds of these latitudes by the lofty range of the Andes, and in part from the monsoon effect, by which the air is drawn in

toward the interior of the continent, and by which they are strengthened but not much changed in direction. At the opposite season of the year this monsoon influence is in the contrary direction, and reverses the directions, giving rise to northerly and northeasterly winds, but of less strength.

Farther south toward the Cape northwesterly winds prevail the whole year, arising in part from the deflection of the prevailing westerly winds of the middle latitudes by the Andes toward the south. The monsoon influence is also here perceptible in causing these winds to be more northerly in winter than in summer.

Along the coast of Peru and on the adjacent ocean, in the latitudes of the southeasterly trade-winds, the temperature of the continent being greater than that of the oceans, there is a tendency of the air the whole year to flow in toward the land, but especially in December, January, and February, which, however, is not strong enough to change the direction, even at this season, of the southeasterly trade-winds so much as to make them southwesterly winds, as in the Gulf of Guinea, but it simply makes them a little more southerly than they otherwise would be.

LAND- AND SEA-BREEZES.

151. Land- and sea-breezes are observed in all countries wherever there is a diurnal alternation of the temperature of the land, above that of the adjacent ocean during the day, and below it during the night—just as there are monsoons where this temperature over a whole continent or large area of country is higher than that of the oceans or surrounding parts of the country during the summer, and below it during the winter. They are observed mostly in equatorial and tropical latitudes, because here the diurnal range of temperature is greatest, and consequently there are the greatest contrasts between land and ocean temperatures; but they are also quite common in middle latitudes, and likewise traces of them were found by Scoresby in Greenland during calm and clear weather. These winds, for reasons already given (§ 132), are light winds

except where there are high lands or mountain ranges near the coast, and extend no great distance from the shore either inland or out into the sea. They depend upon the thermal conditions mostly in the vicinity near the coast, and but little upon those at a great distance toward the interior of continents. They are, therefore, where there are no other causes of disturbance, in a direction perpendicular to the general trend of the coast —inward toward the land during the day, and the reverse at night.

As weak monsoon influences, such as those of Australia, are often not sufficient to reverse the general currents of the atmosphere, such as the trade-winds and the general westerly and southwesterly winds of the middle latitudes, but merely to cause annual oscillations in their strength and direction, so land- and sea-breezes are entirely overwhelmed and not observable where there are other strong prevailing currents, either of the general circulation of the atmosphere, or of monsoons, or other more local and temporary disturbances to be treated farther on, or, if observable at all, it is merely their effect in causing diurnal variations of strength and direction. For instance, if there is a prevailing wind of considerable strength from the ocean inland, the effect of the sea-breeze influence would be superadded to this during the day and increase its strength, while at night the contrary influence would not be sufficient to entirely counteract and reverse it and cause a wind from the land toward the sea, but might diminish very much its strength. Likewise, if the prevailing wind were in the direction of the general trend of the coast, the effect of the land- and sea-breeze influence would be to cause the wind to incline in toward the land during the day and away from it during the night, more or less according to the comparative strength of the disturbing influence. The pure land- and sea-breezes are observed in calm weather only, and of course mostly in clear weather, when the changes of temperature between day and night are the greatest. Along the coasts of India, therefore, and in all monsoon regions, they are only observed for a few weeks at the times of the reversals of the monsoon in spring and fall,

when calms prevail, unless there are prevailing winds from other causes than those of the monsoons.

152. As the effect of the monsoon influence first begins to be seen in the spring as the temperature of the continent first rises above that of the oceans, is the greatest in midsummer, and is reversed in the fall into the winter monsoon as the temperature of the land falls below that of the ocean, and this has its greatest strength in midwinter, so the pure sea-breeze is first felt in the forenoon about 10 o'clock, when the temperature of the adjacent land becomes greater than that of the sea, is greatest at about 3 o'clock in the afternoon, during the warmest part of the day, and is then reversed into a land-breeze about 8 o'clock in the evening, when the temperature of the land becomes greater than that of the ocean; and this has its greatest strength early in the morning, when the temperature is the lowest.

The effect of the land-wind upon the calm smooth ocean in the forenoon is usually observed first near shore, and as the strength of the breeze increases it extends gradually back from the land into the ocean. But the reverse of this, according to Dampier, is often observed. It no doubt depends upon local circumstances. In the case of the true sea-breeze, depending simply upon contrast of temperature between the level coast and the ocean, the greatest thermal, and consequently the greatest pressure, gradient is first caused at and near the coast, where there is an abrupt change of temperature between land and sea; and therefore the interchange of air between sea and land must first commence there, and gradually extend to greater distances, both inland and out into the sea, as the temperature differences increase toward the middle of the day. In the case, however, of the strong sea-winds experienced along coasts where there are steep slopes from high lands and mountain ranges near, it is reasonable to suppose that the interchange may commence first between these slopes and the sea at a distance, and be gradually communicated to the air on the coast, and so the effect would first be seen at a distance, and only afterwards near the coast.

For the same reason that the summer monsoons are stronger than the winter ones in equatorial and tropical latitudes (§ 145), are the sea-breezes here stronger than the land-breezes. Many evidences of this fact might be cited here, and this has been especially observed frequently along the coast of the Gulf of Guinea by many others, as well as somewhat recently by Commander Burke, R. N. He says:[17]

"The winds that prevail along these coasts are already so well known as to render much comment unnecessary, the usual sea- and land-breezes being, with few exceptions, constant; the latter, however, are nowhere very strong, and as a rule are not felt before midnight, nor do they last later than 8 A.M. The sea-breeze commences at about 10 A.M., reaches its maximum strength at 4 P.M., and then gradually falls light; but frequently, and especially in the Gulf of Guinea, it maintains its full strength until it suddenly drops and gives place to the land-breeze."

The strong land- and sea-winds along coasts with long slopes from highlands and mountain ranges depend very much upon the configuration of such coasts. Around high promontories they are comparatively weak, because the area heated above or cooled below the temperature of the sea is small, and the effect is distributed around over a comparatively large area. On the contrary in a narrow bay surrounded by high hills or mountains the effect of the heated or cooled slopes all around is concentrated upon a comparatively small space. The cooled stratum of air at night in contact with the hill-sides flows in from all sides toward the bay with a concentrated effect at the lower part of the bay, or at the common outlet of several valleys which come together, just as floods from the concentrated water of heavy rains increase in volume and force as the streams from all sides come into one large river in the valley. Hence it is said,[18] "The land-breeze frequently comes off in sharp, sometimes in dangerous, squalls; and our Sailing Directions are full of cautions to navigators to look out for the first squalls of the land-breeze: even when it does not come in squalls, it comes quickly."

On the contrary, the sea-winds are strengthened in the same proportion, for when the surfaces of the hill-sides are as

much heated above the temperature of the ocean and the superincumbent air at the same levels above it, the air tends to rush up the slopes and hill-side valleys with the same force as it flows down when they are as much cooled below this temperature, and the air which is drawn up at all sides over a comparatively large area is drawn up the bay and gives rise to a strong wind.

MOUNTAIN AND VALLEY WINDS.

158. On the slopes of mountain ranges and the plains near the base, and especially in deep valleys extending from the slopes into the country below, there are often strong winds, toward and up the mountain side during the day, and the reverse at night. This fact is well known to hunters and all who are accustomed to encamp in the mountains at night; and so the camp-fires are always placed below the tents on the slope, that the smoke may be drawn away by the descending current, instead of being blown toward them. Such currents of course extend to some distance over the plain below, and where the air is brought together by a number of ravines and deep valleys into one common and somewhat contracted current below, the force of the wind may be very great. Just the reverse takes place during the day, for the air then rushes up the mountain side from the plain below, and if this air is concentrated into narrow valleys, either near the base of the mountain or up the side, the current becomes unusually strong.

R. Strachey, Esq., of the Bengal Engineers, in the 21st Vol. Royal Geog. Soc., as given by Espy, says:

"The winds in the mountains of the Himalayas blow up the valley during the day from 9 A.M. to 9 P.M., and down during the corresponding hours of the night. At the debouches of the principal streams into the plains these night-winds blow with great violence, particularly in the winter. They diminish in force as we ascend the mountains, and at great elevations and in the plains of western Thibet the nights are almost perfectly calm. The diurnal winds, on the other hand, in the latter country are terrific; and in travelling there we looked forward to the afternoon, when the winds were at their height, with real dread."

At the time of the total solar eclipse on the 29th of July, 1878, the writer ascended about midday the eastern slope of the mountain to the top of Gray's Peak in Colorado, where the eclipse was total. The day was almost perfectly clear, and the wind above, as indicated by the motion of a few specks of cloud, was westerly. On the way up the mountain there was a very strong current rushing directly up the mountain, such that it was necessary to hold on to one's hat with one hand to keep it from being blown away. This was met by the westerly wind above and by a similar current up the west side, but of less strength because the slope was not so long. On the crest of the mountain, a few rods in width, there was a calm, for the ascending currents from each side after meeting continued on upward, which was indicated by some of the writer's note-paper, on stepping down the east side a few rods, being caught and carried vertically up to a considerable height, and afterwards falling back again near the same spot.

On the descent, late in the afternoon, there was a perfect calm, the mountain slope being now cooled down to about the mean temperature of the day. During the night, especially in the latter part of it, there was, no doubt, a reverse current down the mountain of about the same strength, and continuing until about the middle of the forenoon.

These winds depend upon the same circumstances and are explained upon precisely the same principles as the land- and sea-breezes on mountain slopes along sea-coasts; in fact, the only difference is in the circumstance that in the one case there is ocean and in the other a plain at the base of the slope.

154. A similar but much feebler effect is observed on long declivities of open country with only a very small gradient, such as that between the Mississippi valley and the Rocky Mountains; but in such cases it is generally somewhat masked by other prevailing currents, so that there is not a complete reversion of the wind, but a diurnal change of its direction. The writer has often observed in Missouri and Kansas during clear warm weather, when the prevailing wind is generally southerly or a little west of south, that in the morning when

the surface air is cooled down and tends to flow down the incline, the wind is southwesterly, but as the heat of the day increases it gradually changes around toward the south, and toward evening is from a southeasterly direction, the tendency now being for the heated surface air to move gently up the slope toward the mountains. But during the night, if no abnormal disturbance of any kind takes place, it always gets back again to a southwesterly direction.

CHAPTER VI.

CYCLONES.

155. In preceding chapters we have considered the general circulation of the atmosphere which would arise, upon an earth with a homogeneous and smooth surface, from a regular temperature gradient between the equator and the poles, without regard to differences of temperature in different longitudes on the same parallels of latitude. Such a gradient is deducible from the normals of latitude given in the table of § 68. This gradient, it is seen, varies with the seasons of the year, being greatest in winter and least in summer, and hence there is an annual variation in the general motions of the atmosphere, which has likewise been considered. But upon the earth's surface as it actually exists there are mountain ranges and irregular distributions of land and water, and the effects of these in disturbing the regularity of the general motions by means of deflections and the differences in the frictional resistances on different parts of the earth's surface, and especially between the two hemispheres, have been pointed out.

The temperature gradient from which this general but disturbed circulation of the atmosphere at any time arises, depends upon the difference between the mean diurnal vertical intensity of solar radiation upon the earth's surface between the equatorial and polar regions, which, though variable, never vanishes, and hence the general motions of the atmosphere are variable, but never cease. The general circulation also consists of two systems, which are similar, except so far as they are disturbed on account of a non-homogeneity of the earth's surface, each of which extends over a whole hemisphere.

Subsequently another class of temperature disturbances were considered—those arising from the abnormals of latitude at any time, giving rise to monsoons and land- and sea-breezes. These are fixed in locality, and extend mostly over large areas of the earth's surface, but are subject to annual and diurnal changes, and even to complete reversals with a cessation of motion at the time of reversal, so far as it depends on the monsoon influence, and sensibly so for some little time immediately before and after.

We come now to the consideration of another class of temperature disturbances, which extend over a comparatively small part of the earth's surface and are mostly neither fixed to any given part of the earth's surface, nor do they continue generally for a great length of time, and hence they are of a more local and temporary character than the others. In these there is a gyratory motion of the air around some central point, and hence they are called *cyclones*.

VERTICAL CIRCULATION.

156. If the air over any portion of the earth's surface is warmer at all altitudes than that of the surrounding parts at the same levels, it is lighter, and tends to rise up above its original level, and flow out in all directions above. This decreases the pressure at the earth's surface over this area, but increases it a little over the surrounding parts; and thus there arises a gradient of pressure decreasing from the exterior toward the interior below, which causes a flow of air in from all sides to supply the ascending current. There is thus a vertical circulation and interchange of air between the interior and exterior parts established, just as in the case of the general circulation of the atmosphere (§ 72), except that now the motion is toward the central part below and from it above, while in the general circulation it is toward the polar central region above and from it below, the central region in this case being colder instead of warmer than the surrounding parts.

It is not necessary, however, in order to have such a circu-

lation, that all the strata of air from bottom to top shall be warmer in the interior than in the surrounding parts, but only that there shall be such a disturbance of the equality of temperature that the pressure of any part of the interior is less than that of the exterior part; for where this is the case the air tends to rise up, and that of greater pressure at the same level to flow in and takes its place, and thus the circulation is established. In general the velocity of the ascending current in the interior part is much greater, but over a smaller area, than that of the descending current in the exterior surrounding part over a much greater area, where it is simply a very gradual settling down; for since there is no definite limit, the tendency is for the air above to flow out to still greater distances, and thus to press down and bring in the air below from greater distances.

On account of the non-homogeneity of the earth's surface, comprising hills and valleys, land and water, and dry and marshy areas, all with different radiating and absorbing powers, and also on account of the frequent irregular and varying distribution of clouds, it must often happen that there are considerable local departures of temperature from that of the surrounding parts; and if it should so happen, as it frequently must, that this area is of a somewhat circular form, and the air has a temperature higher than that of the surrounding part of the atmosphere, then we have the conditions required to give rise to a vertical circulation, with an ascending current in the interior, as described above. But unless there is some source of heat by which this interior higher temperature is kept up, this circulation soon ceases, for the interchange of air between the interior and exterior parts of the air comprised in the circulation tends to continually reduce the difference of temperature upon which the circulation depends, and to bring all parts to the same temperature.

157. It has been shown (§ 29) that if any portion of the atmosphere, when it is in the unstable state, receives, from any slight temperature or other disturbance, an upward motion, it becomes warmer than the surrounding air at the same level; and the tendency then is for it to continue to rise as long as the

air is in the unstable state, and thus to give rise to a vertical circulation, and this must continue as long as the atmosphere of the surrounding parts is in the unstable state. Let us consider first the case of a dry atmosphere, and suppose, as in the example previously given (§ 29), that the vertical temperature gradient is that of a decrease of 1.°2 for each 100 meters of increase of altitude. In this case the air of the ascending current becomes warmer than that of the comparatively slow descending current on all sides at the same level at the rate of 0°.2 for each 100 meters of ascent; and so the temperature differences and the force by which the vertical circulation is kept up increase at first until this circulation is fully established and air has ascended from the earth's surface up to the upper strata, after which it gradually decreases; for the interchange of air between the lower and upper strata tends to reduce the temperature gradient of all of it to that of the indifferent state, which is 1° for each 100 meters of difference of altitude, and the horizontal interchange tends to equalize the temperatures between the interior and exterior parts of the circulation, and when this equality takes place all force to keep up the circulation vanishes. Of course the more the unstable state differs from the indifferent state, the greater the energy which maintains the circulation, and the greater the rapidity of this circulation. But in all cases there is probably such an interchange and inversion of the air of the lower and upper strata that the unstable state is soon broken up and the circulation ceases.

In the case supposed above, of a regular and uniform vertical temperature gradient at all altitudes, the differences of temperature at the same levels between the ascending air and that of the surrounding air would increase with increase of altitude; but such gradients are by no means uniform, but may so change at different altitudes that the atmosphere may be in the unstable state below and the stable state above, or *vice versa.* In the first case the ascending air would first become warmer, the differences then decrease, and finally the ascending air become colder than that of the surrounding air at the same lev-

els, but, still the vertical circulation would take place, though its energy would be less, and it would not generally extend up to the top of the atmosphere.

158. In the case of a moist atmosphere with the unstable state for dry air, we have the same energy for originating and maintaining a vertical calculation as in the case of dry air, with the additional energy of all the latent heat of the aqueous vapor set free in its condensation in the ascending current, and this latter is a continuous source of energy as long as moist air is being drawn in from all sides to supply this current. For instance, suppose the air is saturated, and is in the unstable state for dry air, then at the first upward start of the air it becomes warmer than the surrounding air and the tendency is to continue on. As soon as the vertical circulation is fully established, the vertical gradient in this ascending current becomes that given by Table III, which is greater or less according to the season of the year and the altitude. With a summer temperature of 30° at the earth's surface, as may be seen from the example given in § 27, the rate of decrease of temperature is only about 0°.37 for each 100 meters of ascent up to an altitude of 4000 meters, and even up to much greater altitudes the rate would vary but little from this. If the vertical gradient of unstable equilibrium were 1°.2, as assumed in the preceding case, we should have, after vertical circulation was first fully established, a difference of temperature between the air of the ascending current and that of the comparatively quiet surrounding air of nearly 0°.83 for each 100 meters of altitude; for at this time the vertical temperature gradient of the very slowly descending air over a very large area will not have changed much. In the case of dry air it was only 0°.2 for the same vertical temperature gradient; and so of the 0°.83 above, 0°.63 is due to the heat of condensation of the vapor.

If the atmosphere were not completely saturated, but had a depression of the dew-point—air-temperature minus temperature of the dew-point—of 8°, then by Table IV the air at the earth's surface would have to ascend about 1000 meters before condensation would commence, and in this part of the ascent

the rate of the cooling would be 1° instead of $0^\circ.37$ for each 100 meters, after which the latter rate would take place. Hence in this case the whole of the ascending column would have a lower temperature than in the preceding case of complete saturation, being, above the altitude of 1000 meters, equal to $10 \times (1^\circ.0 - 0^\circ.37) = 6^\circ.3$ less than in the case of complete saturation, and hence the energy of the vertical circulation would be much less in consequence of the lack of complete saturation.

In the preceding examples we have assumed a uniform vertical gradient which makes a dry atmosphere unstable. But this is not necessary in the case of a moist atmosphere, in order to produce a vertical circulation. In fact, in the case of complete saturation it is only necessary that this gradient shall be a little greater than that given by Table III, which in the last example is $0^\circ.37$ for each 100 meters. If the gradient were $0^\circ.8$, a little less than that of the indifferent state for dry air, then each part of the air at all altitudes, in its initial ascent of 100 meters, would be $0^\circ.8 - 0^\circ.37 = 0^\circ.43$ warmer than the surrounding undisturbed air at the same levels; and after vertical circulation were fully established, and air had ascended from the earth's surface up to high altitudes, the differences of temperature would be $0^\circ.43$ for each 100 meters up to these altitudes, and so in the upper strata of the air they would be very great. If the vertical gradient were still less but a little greater than $0^\circ.37$, these differences would be smaller, and the force would be able to give rise to and maintain a comparatively feeble circulation only.

159. In the case in which the atmosphere is not completely saturated, we have seen that in the first ascent of the air, until it is cooled, at the rate of 1° for each 100 meters, down to the dew-point, it becomes colder and heavier than the surrounding air, unless the atmosphere is in the unstable state for dry air. It may, therefore, happen that before the air rises up to the altitude where condensation takes place the whole air-column may be heavier than a similar one of the surrounding part of the atmosphere. In this case there must be more than a mere

initial and temporary impulse, or very slight temperature disturbance merely, to start the vertical ascent of the air. The initial temperature of the air before ascent commences must be so much greater than that of the surrounding air, that it does not become cooled down in its first ascent, before condensation takes place, to a lower temperature, on the average for the whole column, than that of the surrounding air.

This will be better understood by tracing the effects which must follow in a few assumed cases. In the following table the first column contains the altitudes of several successive strata at equal intervals, and the column A the corresponding temperatures of the air in its undisturbed state as it existed very nearly on the average of Glaisher's balloon ascensions in clear weather, as given in the table of § 27, assuming that the temperature at the earth's surface was 25° C.:

Altitudes in Meters.		Saturated Air.		$\tau - d = 4°$			$\tau - d = 8°$			$\tau - d = 4°$		
	A	B	C	B′	C′	D	E	F	G	H	I	K
6000	−3°.0	−4°.3	−2°.5	−6°.5	−3°.5	−14°.0	−20°.0	−9°.3	−39°.5	−43°.0	−47°.5	−60°
5500	−1 .5	−2 .7	+0 .3	−5 .0	−0 .7	12 .0	18 .0	6 .9	38 .5	41 .5	42 .9	55
5000	0 .0	−0 .8	3 .0	−3 .3	2 .0	10 .4	15 .5	4 .5	37 .0	39 .5	38 .5	50
4500	+1 .7	+1 .3	5 .5	−1 .4	4 .5	8 .0	13 .0	−1 .9	35 .0	36 .8	34 .0	45
4000	3 .6	3 .3	7 .8	0 .6	4 .8	5 .5	10 .0	+0 .7	32 .5	33 .8	29 .7	40
3500	5 .6	5 .5	10 .1	+2 .8	8 .1	−3 .0	7 .0	3 .2	29 .5	30 .1	25 .4	35
3000	7 .8	8 .1	12 .4	5 .3	10 .4	0 .0	−3 .5	5 .7	26 .0	26 .0	21 .3	30
2500	10 .3	10 .7	14 .6	7 .8	12 .6	+3 .0	0 .0	8 .1	22 .0	21 .7	17 .3	25
2000	12 .8	13 .3	16 .7	10 .4	14 .7	6 .5	+4 .0	10 .4	18 .0	17 .4	13 .6	20
1500	15 .4	15 .9	18 .8	13 .0	16 .8	10 .0	9 .0	12 .7	13 .8	13 .0	10 .0	15
1000	18 .8	19 .0	20 .9	16 .0	18 .9	14 .0	15 .0	15 .0	9 .5	8 .4	6 .6	10
500	21 .0	23 .0	23 .0	20 .0	20 .0	19 .0	20 .0	20 .0	−5 .0	−5 .0	−3 .2	−5
000	25 .0	25 .0	25 .0	25 .0	25 .0	25 .0	+25 .0	+25 .0	0 .0	0 .0	0 .0	0

The columns B and C in the case of saturated air represent—the first the temperatures which the air of each stratum would have after ascending through 500 meters to the stratum above as given in the preceding table; and the second, the temperatures at the different altitudes after the vertical circulation has been fully established and the air has ascended from the earth's surface high up into the upper strata, both being determined by the rates of cooling of ascending air given in Table III. By comparing B and A it is seen that, after this initial start, the temperature of the ascending air is warmer than that of the surrounding and sensibly undisturbed air up to the altitude

of 3500 meters, so that up to that [illegible] the unstable state, and the air being [illegible] any slight impulse or temperature [illegible] to go on; but above the [illegible] small initial ascent, it becomes [illegible] rounding undisturbed air, and [illegible] all directions from the central part [illegible] and a vertical circulation commence[illegible] vertical gradient gradually approximate[illegible] lished current, and the temperature[illegible] umn C. By comparing these temperature[illegible] is seen that the temperatures of the [illegible] than those of the undisturbed air at the same [illegible] far above the altitude of 6000 meters and [illegible] to continue to ascend is then very strong [illegible] decrease of temperatures, therefore [illegible] case of completely saturated air [illegible] clonic disturbance of the atmosphere [illegible] small initial disturbance, and after being [illegible] have considerable violence. But as the air [illegible] in its initial ascent becomes colder and [illegible] rounding air, the vertical circulation may [illegible] before it is deflected off laterally; but [illegible] than the limit of the unstable state on account of the [illegible] tum which it acquires below in its ascent.

If instead of a very small temperature or [illegible] by which initial ascending currents are ind[illegible] area, we suppose the temperature of the interior part of [illegible] over this area to be several degrees higher than that of the surrounding undisturbed parts, then this number of degrees must be added to columns B and C; and then, it is [illegible] the initial force and that after the vertical circulation is [illegible] established are very m[illegible]ncreased as long as this difference of temperature bet[illegible] central and surrounding parts re- main[illegible] verti[illegible]on mi[illegible]iginate in this case if th[illegible] mr[illegible]lien[illegible]n considerably less [illegible]col[illegible]t a[illegible] circulation contin-

ues, this difference of temperature is gradually diminished by the interchange of air between the central and exterior parts, and likewise the general vertical temperature gradient of the whole atmosphere for a long distance all around is gradually brought to that of the indifferent state for saturated air, just as in the case of dry air (§ 157), and then the whole force vanishes.

160. With the same vertical temperature gradient of the atmosphere generally over a large extent of the earth's surface, represented by column A, let us consider the case in which the air is not completely saturated, but in which there is a given difference between the temperature of the air, τ, and that of the dew-point, d, or a depression of the dew-point, $\tau - d$, at all altitudes. In this case the air in its initial disturbance has to rise at all altitudes, by Table IV, about 125 meters for each degree Centigrade of $\tau - d$ before condensation and cloud formation take place, and during this time the rate of cooling is 1° for each 100 meters, the same as in dry air. If we take $\tau - d = 4°$, then the air of each stratum in ascending through 500 meters is cooled down to the dew-point and condensation commences, after which the rate of cooling is that given by Table III, and consequently differs at different altitudes and temperatures. After the air of each stratum from any supposed initial impulse has ascended to the one 500 meters above, the temperatures become those of the column B′, instead of those of A, which represent the temperatures of the surrounding air which has not ascended. By comparing B′ with A it is seen that the ascending air has now become colder, and consequently heavier than the surrounding undisturbed air, and has no tendency to ascend higher, unless the ascent is maintained by the initial starting force, but tends to fall back. It is therefore in the stable state for unsaturated air. If, however, this initial force were continued until the ascending current and the whole vertical circulation were fully established, then the temperatures in such an ascending current would be those of C′ as determined from Table III, allowing a decrease of 5° between the earth's surface and the plane of incipient condensation 500

meters above. It is seen that the temperatures now, from the disengaged latent heat of condensation, are much greater than those before the ascent commenced and those of the surrounding atmosphere generally. If, therefore, the initial temperature disturbance is sufficient to inaugurate a complete vertical circulation, there is then a tendency for it to continue until this tendency vanishes, for reasons already given. If we suppose, for reasons given in § 156, that the temperature of the interior ascending air is 3° greater than that of the surrounding quiet air, we must add 3° to the temperatures of the columns B′ and C′, and then, by comparing these with A, it is seen that not only the tendency of the air to ascend is greatly increased when the vertical circulation is fully established, but that there is this tendency as soon as the air of each stratum has ascended to the plane of incipient condensation 500 meters above, and this tendency after that continues to increase until the circulation is fully established.

161. Let us now assume that the vertical distribution of temperature in the atmosphere over a very large part of the earth's surface is that represented by the column D in the preceding table, and that the depression of the dew-point, $\tau - d$, at all altitudes is 8°. The air of each very thin stratum in this case has to ascend about 1000 meters before condensation of aqueous vapor commences, and in doing so is cooled down 10°. It then has the temperatures of the column E, and consequently is now much colder and heavier than the surrounding undisturbed air, so that if by any temporary impulse it had received such an upward motion, it would not now tend to go on but to fall back, and so in all cases where the vertical temperature gradient is not greater than that of the indifferent state for dry air. But if the initial disturbing force arising from a central higher temperature were sufficient to establish a complete vertical circulation such as to have carried air from the surface up to high altitudes, the distribution of vertical temperature now in the ascending air would be as represented by the column F, the whole ascending air having been warmed up by the setting free of the latent heat in condensation above its

original temperatures of the same altitudes, instead of being cooled below them in the ascent through the first 1000 meters before condensation commenced. The whole of the ascending air now is consequently warmer than that of the surrounding parts, and has a tendency to continue to rise up and to maintain the vertical circulation. Such a circulation, however, could not originate from any slight disturbance of the atmosphere, as in the case of unstable equilibrium of either dry or fully saturated air; but it would require that the ascending air, before ascent, should have a higher initial temperature than that of the surrounding air by about 4° or 5°, in order that, when it has ascended and cooled down to the dew-point, the temperatures may be a little greater than that of the surrounding air, at least up to a considerable altitude, and not have fallen below, and so stopped the further ascent; for if 4° or 5° were added to the column E it would make the temperatures greater than those of the column D up to a considerable altitude, and so the air would continue to ascend and not fall back, and an initial circulation would be established up to at least a high altitude. This additional amount of temperature added to the column F indicates by comparison with D that the force by which the vertical circulation, when once established, is maintained, would be much increased. It is therefore possible, with the preceding assumed vertical temperature gradient and hygrometric condition of the air, for a vertical circulation of the air to originate and be maintained for some time, provided that the initial temperature disturbance is considerable, and not merely such as to give it a slight upward motion, as required in the cases of the unstable state of the atmosphere.

By comparing F with A instead of with D, it is seen that a vertical circulation cannot be maintained with a depression of the dew-point equal to 8° with the same condition of the atmosphere with regard to vertical temperature gradient as in the case in which the depression of the dew-point is only 4°; for in this latter case the vertical circulation is kept up with the vertical temperature gradient corresponding to A, since the temperatures of C′ which exist after complete circulation is

established are greater than those of A, while those of F, in the former case, when complete circulation is established are less than those of A, though greater than those of D. Now the longer the vertical circulation continues, the greater is the depression of the dew-point, for as the moist air ascends and loses its moisture by condensation, and the surrounding air gradually descends, and the cold air of the upper strata with comparatively little aqueous vapor comes down near the earth's surface, the ascending current is supplied with air which is gradually becoming drier, and consequently with air having a lower depression of the dew-point. It follows, therefore, that a vertical circulation which originates from and is maintained by a given vertical temperature gradient gradually loses its force and finally vanishes, because the supply of air to the ascending current becomes gradually drier and its dew-point lower.

As the dew-point becomes lower and the plane of incipient condensation and cloud-formation rises, all the strata in the ascending current below this are reduced to the indifferent state for dry air, and the slowly descending air in the region around is also slowly reduced to this state, so that in time the whole atmosphere in and around the region of ascending air is reduced very nearly down to the indifferent state for dry air, and would be entirely so, if, when the vapor is all nearly condensed, the energy were still sufficient to continue the circulation and it were not broken up by abnormal disturbances before this state is reached.

162. Let us now consider still another temperature condition—one belonging to the winter season with a temperature of 0° at the earth's surface, and let us assume a vertical temperature gradient corresponding to the vertical distribution of temperature represented by the column G in the preceding table, and that the depression of the dew-point is 4°. Proceeding now as before, we get for the temperatures, after the air of the different levels has ascended 500 meters, when condensation commences, those contained in column H; and after complete vertical circulation is established, approximately those of the column I, as accurately as they can be conveniently obtained.

from Table III by an extrapolation for the lower temperatures. By comparing H with G it is seen that when condensation first commences the ascending air at the higher altitudes is colder than the surrounding air, if the ascent does not arise from an initial higher temperature in the ascending air, and that unless this temperature is several degrees greater the tendency is to fall back after the initial ascent of 500 meters. At lower levels, however, the temperature is greater and the tendency is to continue on. If the initial upward motion arises from a single momentary impulse, or only a very small difference of temperature of the ascending air before starting, a vertical circulation at first will take place in the lower strata only, and gradually extend to higher levels; but it cannot extend to very high altitudes, since if such were established we should have the temperatures of the column I, which above the altitude of 5000 meters are less than those of the column G, and consequently of the surrounding undisturbed air at the same levels. The air, therefore, would be heavier and could not continue to ascend above that altitude, but would be deflected from the central part outward in all directions and descend slowly in the surrounding regions toward the earth's surface without disturbing much the air above at high altitudes.

If the air in this case were saturated, the vertical circulation would take place with a smaller vertical gradient, that is, with smaller differences of temperature between the lower and higher strata; but, on the other hand, if the air were drier and the depression of the dew-point were 8° instead of 4°, this vertical gradient would have to be larger, and the more so the drier the air and the greater the depression, until in the case of perfectly dry air, in order to have a vertical circulation, it would have to be greater than that of the indifferent state for dry air, corresponding to the vertical distribution of temperature given in the column K.

The relation between the radiation of dry air and its absorption of solar heat radiation is no doubt such that the unstable state would be induced in a dry atmosphere after a long interval of perfect calm; but in the real atmosphere of nature,

containing more or less of aqueous vapor, we have seen that this state is induced with a much smaller vertical temperature gradient, and before it approximates very nearly to that of the indifferent state of dry air vertical circulations and reversions take place, which tend to bring it back only partially to that of the indifferent state of dry air.

163. In the first assumed cases in the preceding table, in which the air at all altitudes is supposed to be saturated, condensation takes place at once at all levels, as soon as an ascending current is from any cause induced, and the cloud or fog is formed at all altitudes down to the earth's surface. In the second case, with the same vertical gradient of temperature, but with a depression of the dew-point of 4°, incipient condensation takes place at the altitude of 500 meters, and the base of the cloud is at that level; but if the depression of the dew-point were 8°, the base of the cloud would be at the level of 1000 meters above the earth's surface, supposing in all cases that air ascends from the surface. And in general the height of the base of the cloud is about 125 meters for each degree of the depression of the dew-point, varying slightly with different temperatures and altitudes as is seen from Table IV. The drier the air, therefore, the higher are the clouds where the conditions are favorable to the production of ascending currents.

Without ascending currents, therefore, there can be no clouds; and in all the region around that of the ascending air, where there is a gradual descent of air to supply the indraught of these currents, the air is clear; for if even there were complete saturation and cloud from a previous ascent of the air, as it began to descend the air would at once become warmer and unsaturated, and the cloud would soon disappear by evaporation. Of course it must be understood here, as in all that immediately precedes, that the air cools by expansion only. If the air is cooled in other ways, as by nocturnal radiation during a clear night, or by passing from lower to higher latitudes, or from a warmer to a colder surface, as from the ocean to the land in the winter, or *vice versa* in summer, there may be cloud,

called fog, at and near the earth's surface without any ascent of air to higher altitudes to give rise to cooling by expansion.

164. In all of the assumed vertical temperature gradients of the table of § 159, represented by the columns A, D, and G, and the assumed depressions of the dew-point, it is seen, by comparing these columns with those of C, C′, F, and I, that after the ascending current is fully established, its temperature in all of these cases, up to 6000 meters, is greater than that of the surrounding air. It is observed, however, that in the undisturbed atmosphere generally, the gradient becomes much smaller at great altitudes, as is seen in the results obtained from Glaisher's balloon ascents (§ 27), and instead of the gradient being the same at all altitudes, it diminishes with increase of altitude very nearly as the pressure does, and consequently the vertical temperature gradients at very high altitudes become small. They may be, and most probably are, such that the temperature of the ascending air at high altitudes is less and consequently the density greater than that of the surrounding air; and so, even with the momentum which it has acquired in its ascent in the lower strata. it does not ascend to the top of the atmosphere, or even to very high altitudes, but is mostly or entirely deflected horizontally in all directions at a lower level. In such a case the vertical circulation would scarcely, or not at all, extend to high altitudes, but be confined mostly to the lower strata, and the air at very high altitudes be very little disturbed, and this mostly or entirely by means of friction between it and the air currents below.

If the ascending current extends up to a given level only, of course there is no condensation and cloud-formation above that level, and the air remains clear and comparatively quiet above, while below there may be a brisk ascending current and great cloudiness.

165. We have seen in preceding sections, and especially in the examples given in the table of § 159, that the vertical temperature gradient of the ascending air in a cyclone, and the differences between it and that of the surrounding air, may be and must generally be very irregular, so that there is not a

uniform horizontal temperature gradient at all altitudes between the interior and the exterior part, as there is somewhat between the equatorial and polar regions of each hemisphere, the differences of temperature here on different parallels, and consequently the horizontal temperature gradients, being nearly the same at all altitudes. But still the greater temperature in the interior causes an upward expansion of the air and greater vertical distances between the isobaric surfaces here than in the exterior part where the temperature is less, just as in the case of the general motions of the atmosphere these vertical distances are greater in the equatorial than in the polar regions, as explained in § 71. The motions of each part of the air between the interior and the exterior are very similar in the two cases, except that they are reversed. In the general motions of the atmosphere the air moves toward the pole or centre above, descends gradually in the central part, that is, in the polar and middle latitudes, moves from the central toward the exterior part in the lower latitudes and ascends slowly in the exterior or tropical regions; but where the interior is the warmer, the air moves toward the interior below, ascends in the interior part, flows away above toward the exterior part, where it descends very slowly toward the earth's surface, there to commence again another circuit. All that is stated in the former case with regard to neutral planes between the counter horizontal and vertical currents, the satisfying of the condition of continuity, and the explanations of the vertical circulation, holds also in this in a general way. In this, however, there is not that definiteness of outer limit which there is at the equator to each of the hemispheres in the general hemispheric circulation.

GYRATORY MOTION OF CYCLONES.

166. We have seen that while the atmosphere is clear and somewhat quiet, the tendency is for the upper strata to cool down to so low a temperature in comparison with that of the earth's surface and the lower strata, that the unstable state is

induced, if not for dry air, at least in the case of saturated air, and that although the air without ascending currents is rarely, and only near the earth's surface completely, saturated, yet if there is only a slight local temperature disturbance—a little excess of temperature over some central area above that of the surrounding air—a vertical circulation is started which continues until there is such an interchange and inversion of the air of the upper and lower strata that the unstable state is destroyed, and the stable state and comparative calmness and clearness are again restored for a while, until broken up again, in perhaps a short time, in a similar manner as before. If the earth had no rotation on its axis the air in its vertical circulation, in all such cases of local and temporary disturbances, would move in the lower strata directly toward the central part of the area of higher temperature and less pressure and out from it above, as already described, and there would be no gyratory motion around the centre. But in consequence of the earth's rotation, where a body moves in any direction on the earth's surface there is a force which deflects it to the right in the northern hemisphere, and the contrary in the southern (§ 53). In the case of a central force acting upon the body, as where air is forced in from all sides toward a centre by a pressure gradient, this deflecting force is strengthened by the relative gyratory motion of the body itself, as is seen from the expression of F_t, § 52, in which the part depending upon n is due to the earth's rotation, and that upon ν to the gyratory motion; and so very near the centre, where $\nu = v/r$ becomes very large, the deflecting force becomes much greater, with the same velocity of the body s. The air therefore in the lower strata, in being forced in from all sides toward the centre, runs into a gyration around this centre, and this in the northern hemisphere is from right to left, or contrary to the motion of the hands of a watch.

This may be illustrated by means of the behavior of water in a shallow basin, where it is allowed to run out through a hole in the centre. If the basin is at rest, and the water in the basin has no motion of any kind before the hole is opened, the

water flows directly from all sides toward the centre, without any gyratory motion. But if the basin, although the water with reference to the basin is perfectly at rest, has the least gyration around its centre, the water in approaching toward the centre runs into a gyration around that centre, and the gyratory velocity becomes very great near the centre. It is somewhat the same where the air over a considerable area of the earth's surface is forced from all sides toward the centre. This area, we have seen, § 52, gyrates, in consequence of the earth's rotation, around its centre, with an angular velocity proportional to the sine of the latitude; and so the air, in running in from all sides toward the centre, runs into a gyration around the centre, just as the water in the basin. The principal difference in the two cases is, that the water runs down through the hole and disappears, while the air runs upward over the central area, and flows away above.

167. In the flowing away of the air above in all directions from the centre, it is deflected toward the right in the northern hemisphere by the same force as in running in toward the centre. The first effect is to counteract and overcome the gyratory velocity acquired by the air in being forced toward the centre, and after that it produces a gyratory motion in the contrary direction, that is, from left to right. Where the system of vertical circulation and gyratory motions becomes fully established, and the air which flows out above is drawn in again toward the centre below, this gyratory motion from left to right has first to be overcome by the deflecting force before a gyratory motion from right to left begins to be generated. In connection, therefore, with every vertical system of circulation there is produced by the deflecting force of the earth's rotation two kinds of gyrations—the one, mostly in the interior part, from right to left in the northern hemisphere, and the other, mostly in the exterior part, in the contrary direction. The interior, and by far the most violent part, is called a *cyclone*, and therefore the exterior and comparatively gentle part is properly called an *anti-cyclone;* and the two always go together. The gyrations of the former are properly called *cyclonic gyra-*

tions, and those of the other, *anti-cyclone gyrations;* it being understood that each of them is exactly reversed in the other hemisphere.

The deflecting forces upon which the gyrations depend being greatest at the poles and diminishing as the sine of the latitude toward, and vanishing at, the equator, of course the gyrations arising from any given temperature disturbance or vertical circulation, all other circumstances being the same, are most violent in the polar regions, less so with decrease of latitude, and vanish at the equator, the motions there being directly toward and from the centre, the same as would be the case everywhere on the earth's surface if the earth had no rotation on its axis.

168. In the case of no friction between the air and the earth's surface, and the initial state of the air being that of relative rest, each particle of air, in being drawn in toward the centre, would, after the system of motions had become fully established, move in accordance with the principle of the preservation of areas, § 41, provided we consider the absolute gyratory velocity, including both that depending upon the earth's rotation and that which is relative to the earth's surface. Hence, putting w for the absolute gyratory velocity, and r for the distance from the centre, we must at all distances from the centre have the relation, as in § 42,

$$rw = c,$$

in which c is a constant equal to the average value of rw taken while the air is yet at rest, and w has the value rn', for the whole mass of air brought into motion. As r diminishes w must increase, and consequently near the centre it becomes very great, and the value there of w consists mostly of the gyratory velocity relative to the earth's surface, since the part depending upon the earth's rotation decreases as r, and consequently near the centre becomes very small.

If each particle of air, in being acted upon by the centripetal force below, and centrifugal force above, arising from pressure gradients, were entirely free, not only from the effect of the

friction of the earth's surface, but likewise from the effect of other particles acting upon it by contact and by friction, its gyratory velocity would satisfy the preceding expression for all values of r or distances from the centre, with the value of the constant $c = r_0w_0$, in which r_0 and w_0 were the values of r and w before relative motion commenced. But since in the interactions of the different parts of the air action and reaction are equal, these cannot affect the average of rw for the whole mass, and this, therefore, must remain constant and equal to its value before motion commenced, and therefore equal to the average of r_0w_0 for the whole mass. The gyratory velocity at all altitudes is in this case the same at the same distances from the centre, and there is consequently no friction then between the strata arising from relative gyratory velocities. The whole deflecting force is spent in either giving gyratory velocity and momentum to the air, or in overcoming this where already acquired. The air, therefore, has a cyclonic motion at all altitudes in the interior and an anti-cyclonic motion in the exterior part, and the distance from the centre at which the cyclonic changes to the anti-cyclonic motion is the same at all altitudes.

By referring to § 74 it will be seen that this result is very similar to that in the general motions of the atmosphere in the case of no friction, the value of r being the distance from the centre of gyration in this, while in the other it is the distance from the earth's axis of rotation.

169. If the whole region of temperature disturbance and vertical circulation, and the underlying portion of the earth's surface, did not have a gyratory motion around the centre in consequence of the earth's rotation, the vertical circulation would have the full force due to the pressure gradients arising from the differences of temperature between the interior and the exterior, as in the case of the general motions of the atmosphere if the earth had no rotation on its axis, and if there was no friction there would be continued acceleration; but in case of friction the acceleration continues until the friction becomes equal to the forces, after which the circulation

is uniform so long as the disturbing forces are the same. But where the whole gyrates around the centre, we have seen, the deflecting forces tend to produce a cyclonic gyration in the lower strata of the atmosphere, and an anti-cyclonic in the upper strata; but in the case of friction between the earth's surface and the air, and between the different strata having relative velocities, the forces are not entirely spent upon the inertia of the air, as in the case of no friction (§ 168), but partly upon the inertia and partly in overcoming the friction. So far as it is spent upon the inertia, the effect is to cause cyclonic gyrations in the interior and anti-cyclonic gyrations in the exterior, as in the case of no friction between the air and the earth's surface, in which the whole force is spent upon the inertia of the air; but so far as it is spent upon the friction between the different strata, the tendency is to maintain counter gyrations in the lower and upper strata, cyclonic below and anti-cyclonic above. The resultant effect is, therefore, in part the one and in part the other. Whatever the gyratory velocity below may be, that above is less for the cyclonic and greater for the anti-cyclonic at the same distances from the center. The greater the altitude of the stratum, therefore, the nearer the centre does the change take place from the cyclonic to the anti-cyclonic gyrations.

By comparing these results with those obtained in a somewhat different manner with regard to a whole hemisphere in the general circulation of the atmosphere, it will be seen that they are very similar, except that in the general circulation of a hemisphere the gyratory or east components of velocity are greatest above, while in the cyclonic they are least above; and in the general circulation the distance from the pole at which the easterly velocities vanish and change to westerly ones increases with increase of altitude, while in cyclones the distance from the centre at which the gyrations vanish and change from the cyclonic to the anti-cyclonic decreases with increase of altitude. This arises from the difference in the distribution of temperature and of the horizontal temperature gradients; in the case of the hemisphere in the general circu-

lation the polar or central region is the colder, and in the cyclone the central part is the warmer, and consequently the directions of the vertical circulations are reversed in the two cases.

170. Although in the cyclones the gyratory velocities for the whole system, considered algebraically, and regarding the cyclonic as the positive ones, are less above than below, yet there is a limit beyond which the difference cannot go, and this depends upon the temperature gradients between the interior and exterior parts. As soon as gyratory motions are produced there arises another modifying force, the force deflecting to the right of the direction of gyratory motion in the northern hemisphere (§ 52), and so *from* the centre in the case of cyclonic, and *toward* it in that of the anti-cyclonic gyrations. If the gyratory velocities, and consequently the deflecting forces, were the same in the upper as in the lower strata, they would not interfere with the vertical circulation which is maintained by the gradients arising from differences of temperature, as explained in § 63, and the vertical circulation would be maintained by the full force of the pressure gradients arising from the differences of temperature, just as in the case of no gyrations whatever. But if these deflecting forces are greater below than above, it interferes with this vertical circulation, and the difference between the gyratory velocities and the forces below and above may be such as to entirely counteract the forces by which the vertical circulation is maintained, just as forces of different strengths applied in the same direction at the bottom and top of a turning wheel, if the difference is great enough, and the stronger force is applied in the direction which tends to stop it, entirely counteracts the force by which the wheel is kept in motion.

In the case of regular temperature gradients, an expression of the differences of the velocities required for this might be obtained in a function of the temperature gradient, as in the case of the differences of the east components of velocity above and below in the general circulation of the atmosphere (§ 77).

But however irregular the temperature disturbance and the temperature gradients may be, the differences between the gyratory velocities above and at or near the earth's surface may be such as to entirely destroy vertical circulation. But it is evident that the actual differences must fall short of this; for if the gyratory velocities were such as to entirely stop vertical circulation there would then be no deflecting force in a direction at right angles to that of the vertical circulation to overcome the friction between the strata having different gyratory velocities, and so this friction would tend to reduce the velocities above and below to the same, and would at once so decrease the differences between them that the vertical circulation would not be entirely stopped; but still its velocity would be comparatively much smaller than in the case of no gyration of the whole system around a centre, in which cases the vertical circulation is not hindered by differences in the deflecting forces below and above, but the vertical circulation is maintained by the full force of the pressure gradients arising from the difference of temperature between the interior and the exterior. These deflecting forces, therefore, act as a sort of governor, as in the case of the general motions of the atmosphere (§ 75). If the differences between the gyratory velocities below and above, and consequently between the deflecting forces, are a little too great, the velocity of the vertical circulation becomes too small, and the effect of this is to diminish these differences, and *vice versa*.

Since the differences of the velocities above and at the earth's surface, in order to counteract the vertical circulation, must be such that the differences of the deflecting forces above and at the earth's surface are adequate to counteract the forces by which the vertical circulation is maintained, and these deflecting forces at the same distance from the centre are proportional to the temperature gradient there between the interior and exterior part of the cyclonic area, it is evident that the greater the temperature gradients, the greater the gyratory velocities above differ from those at or near the earth's surface, and very nearly or quite in proportion, since the limits which

they do not reach, but fall short of a very little, are in that proportion.

But the deflecting forces, as seen from the expression of F_v, § 52, for the same gyratory velocity v, increase as r decreases, and near the centre very nearly inversely as r, so that the relative velocities of the strata below and above, required to counteract the effect of any given temperature gradient, are smaller the nearer the centre.

171. The conditions of temperature and of temperature gradients determine the relations between the gyratory velocities above and those near the earth's surface, whatever the latter may be. These depend upon the amount of frictional resistance of the earth's surface to the gyrations and upon the forces which overcome these resistances, and must be so adjusted that the principle will be satisfied that all the resistances from every differential element of the earth's surface, so far as the gyrations extend, multiplied into the distances from the centre, or, in other words, the sum of all the moments of couple must be 0, or else the tendency would be to turn the earth around the axis the pole of which is the centre of the cyclone. But this cannot arise from central forces alone, either centripetal or centrifugal, such as arise from the temperature gradients, and these are the only real forces concerned, the deflecting forces, so called, being simply modifications of the directions of motion as referred to the gyrating surface of the earth.

From the centripetal motion of the air in the lower part of the atmosphere arises the deflecting force which gives rise to a cyclonic motion, and from the centrifugal motion in the upper strata, that which counteracts the cyclonic and tends to produce an anti-cyclonic gyration. Since there must be as much motion toward the central part below as away from it above, with a slight exception while the cyclone is forming or vanishing, in order to satisfy the condition of continuity, these deflecting forces are exactly equal and contrary to each other, and so, taken together, have no tendency to overcome the friction between the air and the earth's surface in the cyclonic and anti-cyclonic gyrations, but tend mostly, after the vertical

and gyratory circulations are fully established, to overcome the friction between the strata of air having different gyratory velocities, and in part to change the gyratory moments. Before that, of course, especially at first, the forces are mostly spent in overcoming the inertia of the air in producing or destroying these moments. As the air which moves in toward the centre below has a greater gyratory velocity than it has at the same distance from the centre in flowing out above, the air, from the time it enters below within a given distance of the centre until it passes out above beyond that distance, loses a certain amount of kinetic energy corresponding to the difference of gyratory velocities at the times of entering within, and passing without, this limit; and this amount, meanwhile, has been spent upon the frictional resistance of the earth's surface within the given distance from the centre. A part of the deflecting force above is spent in decreasing the gyratory velocity there, and consequently only what is left counteracts, through friction between the strata, the equal deflecting force in the contrary way below, and the remainder not counteracted is left to overcome the friction at the earth's surface, but this is equal to that arising from the gradual loss of the kinetic energy.

We have just seen that the relative velocities of the gyrations above and below decrease directly as the radial temperature gradients, and also decrease with the decrease of r, the distance from the centre, but not in proportion. But the temperature gradient vanishes at the centre, and must be very small for some distance near the centre. The amount of momentum lost, therefore, as the air ascends to higher altitudes, and consequently the force upon which the gyratory velocity at the earth's surface is maintained, is very small near the centre, and consequently the gyratory velocity there is comparatively small, where there are frictional resistances to be overcome, though where there are no such resistances they become greater very nearly inversely as the distance from the centre.

In the exterior part of the cyclonic system, where the gyrations at the earth's surface are anti-cyclonic, it is readily seen

how the air in the upper strata, where it has a greater anti-cyclonic gyratory velocity, in settling down gradually into the lower strata in its vertical circulation, where the gyratory velocity is less, gradually loses its momentum, and that this lost momentum, by means of friction, is transferred to the surface, where it becomes a force applied directly to overcoming the frictional resistance to the anti-cyclonic gyrations.

172. The gyratory velocities at the earth's surface corresponding to the frictional resistances which are equal to the forces overcoming them, of course depend upon the nature of the surface and the law of resistance with regard to different velocities. For a homogeneous surface, however, whatever this law, it is evident that the velocities of the cyclonic gyrations must, upon the whole, be much greater than those of the anti-cyclonic, in order to have the sums of the moments of couple equal, since the former are much nearer the centre, unless the cyclonic gyrations cover a much larger area. But this is not the case, but rather the contrary; for the whole cyclonic area having no definite limit, the anti-cyclonic part is spread over a much larger area, and therefore the gyratory velocities are very small in comparison with those of the interior cyclonic part.

The less the amount of frictional resistance at the earth's surface corresponding to any given gyratory velocity, the greater the gyratory velocities, so that at sea, with a smooth surface, the gyrations, with the same amount of energy spent, are greater than on land. And in the case of no such frictional resistances, as we have seen, they would become very great near the centre, the principle of the preservation of areas being satisfied in this case, and the whole deflecting force in this case would act in overcoming the inertia of the air and producing gyratory momentum as it proceeds toward the centre in the cyclonic part, and in overcoming this momentum as the air flows back from the centre.

ATMOSPHERIC PRESSURE IN CYCLONES.

173. Whatever the nature of the earth's surface and the gyratory velocities there, it has been shown that the differences

between these velocities and those at any altitude above at the same distance from the centre are the same in any case, so that where the gyratory velocities at the earth's surface are greater for any reason, all those at any altitudes above are increased by the same amount. This however must be taken in an algebraic sense toward the outer limit of the cyclonic part at the earth's surface, where the cyclonic gyrations below change to the anti-cyclonic above; that is, an increase of the cyclonic velocities below diminishes the anti-cyclonic gyratory velocities above. If we imagine the resistances at the earth's surface to be so great that there are sensibly no gyratory velocities there, then the same differences will exist between these and the gyratory velocities above, now all anti-cyclonic, and these are such as to approximate very nearly to that limit at which, by means of the deflecting force toward the centre, the air, expanded upward by a higher temperature in the interior than the exterior part of the whole area of cyclonic disturbance, is completely hindered from flowing away above from the interior to the exterior part. It has been explained that the differences between the gyratory velocities below and above must fall a little short of this limit so as to allow a little vertical circulation, and consequently small gradient of decreasing pressure toward the centre at the earth's surface to keep up this circulation, as explained in § 63. The pressure, therefore, at the earth's surface is scarcely affected by the upward expansion of air in the interior from a higher temperature, since the anti-cyclonic gyrations almost entirely prevent its flowing away above, and the consequent diminution of the pressure at the earth's surface.

All the other conditions remaining the same, if we now suppose that there are gyratory velocities at the earth's surface, then all the velocities at all altitudes are changed by the same amount, taken algebraically when the gyrations change with change of altitude from positive to negative, that is, from cyclonic to anti-cyclonic, or the reverse. Hence the pressure gradients change by the same amount proportionally at all altitudes, and the pressure gradient at the earth's surface, depends

simply upon the gyratory velocity at the surface, with the exception of the small gradient referred to above, sufficient to overcome the friction of the centripetal flow below.

The method of obtaining the barometric gradient corresponding to any given gyratory velocity by means of Table V, has been explained in § 57. If, in connection with the gyratory velocity, there is a component of velocity toward or from the centre, then the gradient required to maintain this velocity, must be added.

Since the barometric gradient is proportional to the horizontal force which causes it, and this, as we have seen, depends almost entirely upon the gyratory velocities, then it follows from the expression of this force, F_v in § 52, that near the centre of the cyclone, where r is small, and consequently $\nu = v/r$ is large, that this gradient is very steep in comparison with what it is for the same gyratory velocity v at a distance from the centre, where r is comparatively large. But v also is largest somewhere near the centre, and consequently the gradients here are comparatively very steep.

For instance, suppose on the parallel of 40° we had in the anti-cyclonic part a gyratory velocity of 20 meters per second at the distance of 2000 kilometers. From Table V we have $0.1010 \times 20 = 2.02$ for the gradient of pressure increasing toward the centre. But this must be *decreased* by the centrifugal force of curvature, since here the two forces are in contrary directions, in the ratio of $v/r = 20/2{,}000{,}000 = 0.00001$ to $2\,n \sin l = 0.0000917$, by Table V, or 1/9 very nearly. In the cyclonic part, at the distance of 200 kilometers, the part of the gradient of pressure increasing in a direction from the centre due to the earth's rotation is the same as before, but now the centrifugal force is 10 times as much as in the other case, and is in the same direction as the other. We therefore have in this case the gradient $G = 2.02 + 2.02 \times 10/9 = 4.26$. Still nearer the centre, with the same gyratory velocity, the gradient, it is readily seen, would be still much steeper. The gradient therefore, in the anti-cyclonic part, even in this case, is small in comparison with what it is near the centre, but the gyratory velocity

in the anti-cyclonic part, we have seen, is also much smaller, and so for both these reasons it is always comparatively small here.

174. Since the deflecting forces of the cyclonic and the anti-cyclonic gyrations are *from* the centre in the former, and *toward* it in the latter, the greatest pressure is at the distance from the centre where they vanish and change from the one to the other, at least so far as the pressure and pressure gradients depend upon these forces. The difference of pressure between the lowest at the centre and the highest where the gyrations vanish, depends of course upon the summation of the pressure gradients, and as these are comparatively steep in the cyclonic part, the difference between the highest pressure and that at the centre is generally very much greater than that in the anti-cyclonic part between the highest pressure and that of the general undisturbed surrounding pressure.

What has just been stated with regard to pressures at the earth's surface is true of those at any altitude above the surface. For the atmosphere above any given level can be regarded as a separate atmosphere, and the gyratory velocities of the general atmosphere at that level as those of the base of the atmosphere above that level. At any given altitude, therefore, the highest pressure in the plane of that altitude is also, or very nearly, where the gyrations vanish and the cyclonic change into the anti-cyclonic. But the greater the altitude the nearer the centre, as we have seen, does this change take place, and the conditions may be such that at considerable altitudes the gyrations may be anti-cyclonic at all distances from the centre, and in this case there is no central area of low pressure, but the greatest pressure is in the centre, and the gyrations are all anti-cyclonic; and considering the part of the atmosphere above this level, we have here an anti-cyclone alone, with no interior cyclone, while at lower levels, the gyrations are partly cyclonic, and partly anti-cyclonic, the former increasing and the latter decreasing in area, and the barometric depression in the centre becoming deeper as the altitude is diminished, until we reach the earth's surface.

All that area in a cyclone in which the barometric pressure is below the usual average, say 30 inches or 760 mm., is called an *area of low pressure*. It is evident that the depth of the minimum pressure depends both upon the extent of the area and the gradients at the different distances from the centre. Where the gyratory velocity is very great near the centre, the gradient depends almost entirely upon the centrifugal force, especially in low latitudes, and as this becomes very great where the radius is small, such cyclones, even if of small extent, may have a deeper minimum pressure than large cyclones.

RESULTANT MOTIONS.

175. Since the motions of the air in a cyclone consist of both a vertical and gyratory circulation, the resultant at any place depends upon both of these motions as rectangular components, and hence the direction of motion, upon the relations of these components. In the lower part of the atmosphere the horizontal motion of the vertical circulation of the air is mostly from the exterior toward the centre, and hence here in both the cyclonic and anti-cyclonic parts of the system, the resultant direction inclines from the tangent of gyratory motion in toward the centre, and the angle between this tangent and the resultant direction is called *the inclination*. But in the upper part of the atmosphere, above the neutral plane, where there is no interchanging motion between the interior and the exterior of the area of cyclonic disturbance, the motion is outward, and hence the resultant direction deviates toward the exterior from the tangent, and the inclination in this case is negative or outward. In the neutral plane, of course, there is no inclination, and the only component of motion here is that of gyratory motion.

Since the horizontal component of the motion of the air, in its vertical circulation, is gradually retarded as the air approaches the centre of the cyclone, and becomes 0 at the centre, and small even at a considerable distance from the centre, while the gyratory component of motion here is usually large,

the inclination of the resultant or cyclonic motion near the centre is small in comparison with what it is in the outer part of the area of violence and of the cyclone proper, and so much more nearly circular.

At and near the earth's surface, just beyond the ring of highest pressure, there is an exception to the inclination toward the centre. In consequence of the anti-cyclonic gyratory motion being retarded here by the greater amount of friction, the deflecting force toward the centre here is so much weakened that it is not equal to that of the pressure gradient, and the air, instead of flowing in toward the centre, is forced out in the contrary direction from beneath the high pressure. Hence here the resultant direction is anti-cyclonic and outward and the inclination negative, as in the upper part of the atmosphere.

On the other side of the ring of highest pressure the tendency, for the same reason, is for the air to be forced out from beneath toward the centre, and consequently this force combines with that of the general gradient near the surface arising from differences of temperature, upon which the vertical circulation depends, and so increases this component of motion very much, and causes the inclination here to be much greater than it otherwise would be, and also greater than that in the higher strata immediately above.

On both sides of the ring of highest pressure the gyrations are strengthened by the outflow of air beneath on each side, for the deflecting force depending upon the earth's rotation arising from this outflow is in both cases in the right direction for this. In fact, on the outside of this ring there could be no anti-cyclonic gyration if it were not for this force.

176. We have seen that if it were not for the viscosity of the air, the relative gyratory velocities below and above would be such that all vertical circulation would be prevented, and so in this case all the conditions of the problem would be satisfied with a gyratory motion simply, whatever these may be at the earth's surface. But with the least viscosity, the tendency is to reduce these relativeve locities to 0, unless there is

a little vertical circulation sufficient to give rise to deflecting forces, in the one direction below and the contrary above, sufficient to overcome the friction which tends to diminish these relative velocities. The less the friction, therefore, due to viscosity, between strata of air moving with different velocities, the less of the vertical circulation is required. The less the friction, therefore, the less the inclination of the resultant motion, this entirely vanishing where there is no friction. Hence near the earth's surface, where there is much friction, there is a larger inclination, while in the upper strata it is comparatively small and the gyrations are more nearly circular.

It must not be understood, however, that this is true in general for all parts of the system, but only at any given place; for the inclination depends upon the condition of continuity and other conditions. For instance, near the centre there is a tendency of the air to run into rapid gyrations, while the veloc ity of the inflow of air toward a centre, or any barrier by which it is stopped, must gradually become less and less and finally vanish. Near the centre of a cyclone, therefore, the ratio between the radial and the gyratory velocity, and consequently the inclination, is in general less than at greater distances.

SURFACE CALMS.

177. It has been shown (§ 171), that near the centre of a cyclone the force which tends to keep up a gyratory motion becomes very small. There is, therefore, little or no motion of that sort for some distance from the centre. But we have just seen above that the motion to or from the centre also becomes small near, and vanishes at, the centre. However violent the gyrations of a cyclone, therefore, may be at some distance from the centre, at and near the centre there is always sensibly a calm, extending to greater or less distances, according to the dimensions of the cyclone and other circumstances.

Under the ring of highest pressure there is no gyratory motion, for we have seen that the gyratory velocities must necessarily vanish and change sign at some distance from the

centre, and it has been shown that hère is the place where there is the highest pressure. And as the air also flows out from beneath this highest pressure, on the one hand toward the interior and on the other toward the exterior, it is evident that there is here no radial motion. There being, therefore, neither gyratory nor radial motion, there must be here a ring of calms.

But on account of the very small gyratory velocities (§ 174) and consequent pressure gradients at the earth's surface in the anti-cyclone, as compared with those of the cyclone, and also because here the radial velocity depends upon the differences between two forces, the one arising from the temperature and resulting pressure gradient, which tends to keep up the vertical circulation, and the other arising from the tendency to flow out from beneath the high pressure, the motions of the air at the earth's surface in the anti-cyclone are very gentle, and generally of about the same order as the various abnormal disturbances and irregularities, so that they are not readily distinguished by observation; but that such anti-cyclonic and outward motions do exist, and a corresponding pressure gradient, with a maximum pressure at some intervening distance between the centre and exterior limit, is without doubt.

GRAPHIC REPRESENTATION OF MOTION AND PRESSURE.

178. In Fig. 1 is given a graphic representation of the resultant motions and of the barometric pressures for both the surface of the earth and for some level high up in the atmosphere and above the neutral plane, where the motions in the vertical circulation are outward from the centre. The solid circles represent isobars at the earth's surface and the solid arrows the directions, and in some measure, by their different lengths, the relative velocities of the wind. The heavy circle represents the circle of greatest barometric pressure at the earth's surface, say 765 mm., while the pressure of the outer border is 760 mm., and the dividing line between the cyclonic and anti-cyclonic gyrations. Within this limit the pressure diminishes to the centre and the gyrations are cyclonic and

the direction of the resultant of motion inclines in toward the centre, but beyond that limit the gyrations are anti-cyclonic and the direction of resultant motion inclines toward the outer border of these gyrations. The heavy dotted circle represents the circle of maximum pressure at some high level, and is much nearer the centre than that at the earth's surface. It is also

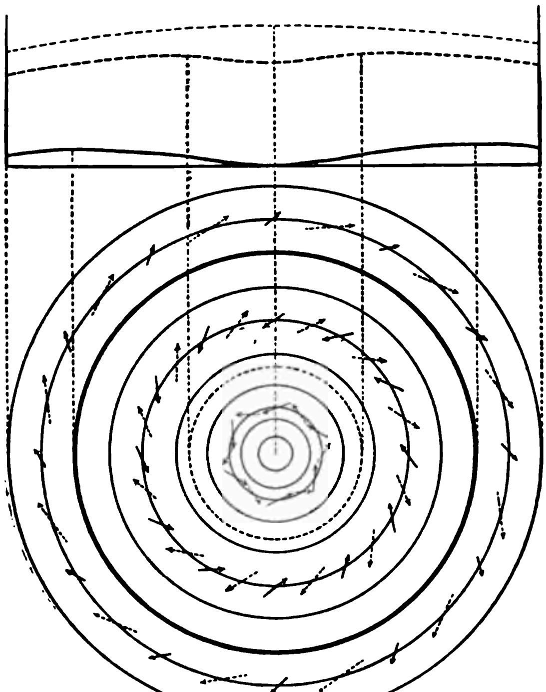


Fig. 1.

the dividing line between the cyclonic and anti-cyclonic gyrations at that level. The dotted arrows indicate the directions, and in some measure the relative velocities, of the wind at this level. The arrows in the cyclonic part represent the direction of the wind as declining outward, because the plane here considered is supposed to be above the neutral plane, where the radial component of motion is outward, but for any level be-

low the neutral plane the inclination is still inward. The arrows are shorter above in the cyclonic part and longer in the anti-cyclonic part than they are at the earth's surface, since the cyclonic gyratory velocities decrease and the anti-cylonic increase with increase of altitude.

The upper part of the figure is a representation of a vertical section of the air, very much exaggerated in altitude, in which the solid curved line represents a section of an isobaric surface near the earth's surface, say of 740 mm. barometric pressure. The lowest part corresponds with the centre of the cyclone and the highest part with the heavy circle in the lower part of the figure, and the steepest gradients with the longest solid arrows, since the greater the gyratory velocities at the earth's surface the greater the gradients, though they are not strictly proportional. The second dotted curved line from the top represents a section of the isobaric surface of high altitudes, in which the highest parts correspond with the heavy dotted circle below, since the highest pressure at all altitudes is very nearly where the cyclonic gyrations vanish and change to the anti-cyclonic. The depression here is smaller because the cyclonic area is smaller, and the gyratory velocities less, than at the earth's surface. The upper dotted line belongs to an isobaric surface still higher, where the gyrations are supposed to be all anti-cyclonic, and here, consequently, the greatest pressure is in the centre, as indicated by the curved line.

As the interior of the whole cyclonic system is warmer than the exterior, and consequently the air less dense, the distances between the isobaric surfaces are necessarily greater in the interior than the exterior part, and so, however much the isobaric surface at or near the earth's surface may be depressed by the cyclonic gyrations there, at a considerable altitude, if the temperature difference is great enough, it must become convex instead of concave.

The track of any given particle of air in a cyclone, resulting from the vertical and gyratory circulation, is that of a large converging and ascending spiral in the lower part, but of a diverging and ascending spiral in the upper strata of the at-

mosphere, and the nearer the earth's surface the more nearly horizontal is the motion, since the vertical component gradually decreases and vanishes at the surface.

The whole energy of the system by which the inertia of the air and the frictional resistances are overcome and the motions maintained, is in the greater interior temperature and the temperature gradients, by which the circulation is maintained. This being kept up, the deflections and gyrations are merely the result of the modifying influence of the earth's rotation, which is not a real force, since it does not give rise to kinetic energy, but merely to changes of direction.

It must be borne in mind that the preceding is a representation of the motions and pressures of a cyclone resulting from perfectly regular conditions, in an amosphere otherwise undisturbed, and having a uniform temperature, except so far as it is affected by the temperature disturbance arising from the cyclonic conditions. Accordingly results so regular are not to be found in nature, but generally only rough approximations to them.

Since the wind inclines less and less toward the centre of the cyclone below the neutral plane and declines from the centre above it, the upper currents above this plane in a cyclone are always from a direction, in the northern hemisphere, a little to the right of that of the lower currents, when not affected by abnormal circumstances.

COMPARISONS WITH OBSERVATIONS.

179. All the phenomena, as deduced from theoretical considerations, have been confirmed by observation. Areas of low pressure with encircling winds, properly called cyclonic depressions, are constantly occurring in nearly all parts of the world, and their delineations on the charts of the different Weather Bureaus are familiar to many persons. A remarkable example of this kind has been recently given by Mr. Harding[1] of the storm which swept across the British Islands on December 8th and 9th, 1886, and which was one of the most violent dis-

turbance swhich had occurred for many years. The following figure is a copy of his chart given for 6 P.M. of the 8th so far as observations were available. From this it is seen that the barometric depression in the centre is more than two inches, and that the gradients, especially on the southeasterly side, are very steep, indicating very strong winds. The arrows indicate that the motion of the air is cyclonic around the lowest pressure with an inclination toward the centre.

On the southeasterly side the difference of the barometer was 1 inch in 240 miles, giving a gradient of about 7 mm. The greatest observed velocities were over 70 miles per hour, giving a gyratory velocity a little less of about 30 m. per second, and these most probably corresponded very nearly with the preceding gradient. A straight-lined wind on the parallel of 50° with a velocity of 30 meters per second, by Table V, would give a gradient $G = 0.1204 \times 30 = 3.6$ mm., and consequently would not nearly give the observed gradient of 7 mm. But if we suppose the gyration had a radius of curvature of 400 kilometers, then, as explained in § 173, this gradient is increased from this cause by a quantity which is to it in the ratio of 30/400000 to $2\,n\,\sin\,l = .00009$, by Table V, and so by a quantity equal 3.0 mm. The gradient, therefore, with this radius of curvature becomes 6.6 mm., very nearly equal to the observed gradient. But to this a small unknown addition has to be made for the pressure gradient resulting directly from the temperature gradient.

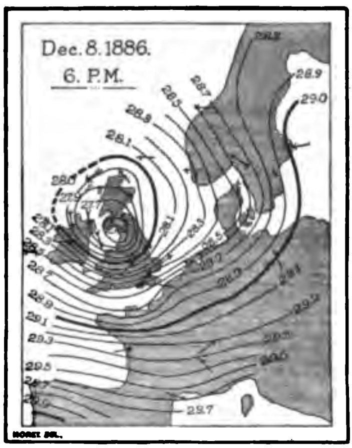


Fig. 2.

180. The inclination of the winds at the earth's surface in a cyclone was observed by Redfield in the great cyclones which originate within the tropics and progress northward and north-

eastward to the higher latitudes, but no estimates were made of the average angle of inclination.

Soon after this Dr. Buys Ballot gave his law, of more general application, namely, that winds always blow, in the northern hemisphere, with high barometer to the right and low barometer to the left of the direction in which they blow, but that this direction inclines a little toward the area of low barometer. Mr. Ley was the first to determine the average angle of inclination in large cyclones from a large number of observations made in the British Islands and the adjacent parts of the continent.[39] His results were, 1°, that the winds commonly incline from the districts of higher toward those of lower pressure; 2°, that this inclination is much greater at inland than at well exposed sea-stations. The collective mean result for 15 stations was an angle of inclination of 20° 31′. The mean for five of the well exposed sea-stations was 12° 49′, while for five of the most inland stations it was 28° 53′. This is in accordance with the theory, by which the angle is increased by increase of friction at the earth's surface. He also obtained a much larger inclination on the east than on the west side of the cyclones.

Subsequently Mr. Ley determined the relation between the direction of both the under and upper currents of the air in areas of low pressure from a much larger number of observations in the British Islands, France, Spain, Switzerland, Austria, Turkey, Russia, Denmark, Sweden and Norway.[40]

His table of results is so important, not only for our present purpose, for which only those results belonging to the surface are needed, but likewise in comparisons further on (§ 205), that we shall give it here in full, though in a different form and different notation. The following table and accompanying figure, taken from the Meteorological Researches, Part II, Coast Survey Report for 1878, give his numerical results. Each area of low pressure was divided into 16 districts as represented in Fig. 3. Considering for the present the results only of the surface winds, we find that the average angle with the radius for the exterior districts, denoted by the large letters, is 64°.6, and for the interior districts, denoted by the smaller letters, 65°.9.

	Surface Winds.		Upper Currents.	
Districts.	No. of observations.	Mean angle with radius.	No. of observations.	Mean angle with radius.
A	198	62°	51	—5°
B	407	52	173	163
C	511	48	226	152
D	675	54	290	146
E	803	66	328	124
F	378	76	199	101
G	277	79	81	96
H	196	80	43	99
a	195	65	58	172
b	391	53	104	130
c	426	58	94	135
d	454	55	141	102
e	629	64	135	73
f	402	74	142	51
g	250	77	83	90
h	204	81	46	106

This gives an average inclination of about 25°, a little greater than before. But he now has mostly inland stations and only

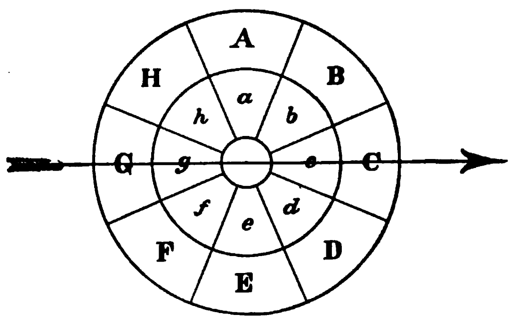


Fig. 3.

few coast stations in comparison, and so this is also a confirmation of theory.

From the Weather Maps of the United States Signal Service for the years 1872 and 1873 Professor Loomis obtained an average inclination of nearly 47°, which is much greater than that obtained by Mr. Ley for Europe. These two results, however, are not inconsistent with each other. The difference may be in a small measure due to difference of latitude, since theory

requires greater inclinations for lower than for higher latitudes, but it is no doubt due mostly to the fact that Loomis took in all observations within isobars of 29.9 inches, and hence many cases of only very small velocities, while Mr. Ley took in cases mostly of the more violent winds nearer the centres of the cyclones. The winds from the outer border of the cyclone proper, where the motion of the air is more toward the centre and has not yet assumed much gyratory velocity, of course, have a greater inclination than those nearer the centre, where the gyratory velocity is greatest, and where, it has been shown, the inclination must be less (§ 175).

In subsequent researches with a greater number of observations and measurements taken from the Signal Service Weather Maps, Loomis[41] deduced an average inclination a few degrees less than that given above. From this discussion it resulted that the inclinations at smaller distances from the centre are less than at greater distances. This is, again, in accordance with theory. The inclinations ranged from about 37° for an average distance of 250 kilometers to 44° at the average distance of 1,200 kilometers.

The average inclination which he obtained at the same time for low barometric pressures from Hoffmeyer's charts of the northern part of the North Atlantic ocean is considerably less than in the preceding case, as it should be in the case of a water surface, where the friction is less. The inclination in this case likewise increased with increase of distance from the centre, being 25°.4 at the distance of 278 kilometers, and 34°.8 at the distance of 1535 kilometers. For barometric pressures above 760 mm. the inclinations were still larger.

The late J. Allen Brown from the observations of Makerstown, Dublin and Greenwich, 1843–1848, obtained the following results:—[44]

	Jan.	Feb.	Mar.	April	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Year
$\theta =$	257°	293	267	260	22	262	274	287	284	267	240	255	269
$\phi =$	247	280	239	260	0	240	249	264	243	245	226	257	249
$\theta - \phi =$	20	13	28	0	22	22	25	23	41	22	14	−2	20

in which, reckoned from N. around by E ,

$\theta =$ the direction of the isobars.

$\phi =$ the direction of the wind.

The deviation, therefore, of the direction of the wind from that of the isobar is 20° to the left on the average of the year.

The following contains the average directions of the cirri, ψ (day only), and the average directions of the surface winds ϕ, made day and night:

	Jan.	Feb.	Mar.	April	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Year
ψ =	279°	303	301	242	291	263	269	267	274	276	236	308	277
ϕ =	241	287	244	274	24	233	230	247	244	237	211	249	243
$\psi-\phi$ =	38	16	57	−32	−94	30	39	20	30	39	25	59	34

These results indicate that for the average of the year the upper currents come from a direction 34° to the right of the surface winds, and that they come from the right in the case of each separate month except two. Supposing the winds to be mostly cyclonic, this confirms the theoretical deduction of § 178.

Mr. Brown also stated that M. Quetelet observed the directions of the cloud motions (without distinction of species) at the Brussels observatory during the years 1833 to 1846, and found the resultant direction by Lambert's formula to be for the 14 years $\psi = 257° 50'$. Also that by noting the difference of directions of surface currents and different kinds of clouds, he obtained from the averages of large numbers of observations,

Cirrus current	—Surface currents, or	$\psi-\phi=29°.6$
Cirro-stratus current—	" "	" $=22.8$
Cumulus current	— " "	" $=14.5$

This is likewise a confirmation of the deduction that high currents come from a direction to the right of the lower currents and the higher the more so.

The observations of Padre Viñes[2] on the hurricanes of the Antilles also show that there is an inclination of the winds toward the centre, and that this is least nearest the centre. In all these hurricanes it was observed that " the gyrating winds cease to be circular at a long distance from the vortex, and are found to deviate from the tangent to the circle with an inclination toward the centre, forming a kind of large converging spiral." This converging is likewise said to " vary, not only in different hurricanes, but likewise in the same hurricane with different directions and intensities of the wind, and with different distances

from the vortex." In the same connection it is also stated that it is "especially small at no great distance from the vortex."

In more recent researches Hilderbrandsson[43] has deduced from observations at Upsala, including those for both high and low pressures, an average inclination of 40°.2, and for the average of the three maritime stations Wäderobed, Utklippau, and Landöu, 32°.2. He likewise obtained for these three stations, as Ley did for the British Isles,[39] Hoffmeyer in Denmark,[44] and Spindler in Russia,[45] a larger angle of inclination on the east than on the west side of the cyclone. This angle also increased with increase of distance from the centre, though the differences near the centre were very small.

The following is the summary of the practical results obtained by Captain Toynbee from a discussion of the observations of the North Atlantic Ocean during the great cyclone of the 24th and 25th of August, 1873:

1. There is strong evidence that the wind in a hurricane draws in towards its centre.

2. The indraught is probably greater in one quarter than another.

3. The indraught is probably greater near the centre than farther from it.

The average inclination from all the observations was 29°, the average latitude being about the parallel of 50°.

Piddington gives as an effect, and also a proof, of the inclination of the winds toward the centre, that ships are often surrounded or have their decks covered, during the passage of the calm centre of cyclones in the neighborhood of land, with land and aquatic birds, butterflies, horseflies, etc. He says:

> "They must be carried inwards by the incurving of the winds. and at the centre are kept there because they cannot fly out of it, and when the ship reaches it of course they make a last effort to reach her as a resting place."

181. From the deductions of theory confirmed and supplemented by numerous observations, it is evident that many of the usual rules and sailing directions must be very much modi-

fied, especially in low latitudes. Although the horn-cards of Piddington, and all rules based upon the strictly circular theory of the winds, may be still used at sea in high latitudes without great error, yet nearer the equator they must become more erroneous, and almost entirely fail very near the equator. It is beginning to be pretty generally acknowledged that in sailing directions in a storm, some allowance should be made for a certain small amount of inclining of the winds toward the centre, but it seems to be thought that this should be the same at all latitudes and at all distances from the centre of the storm. If in low latitudes the inclination of the winds according to theory may be 60° or more, the usual rules for the determination of the direction of the dangerous centre of the storm based upon the circular theory would lead to an error of five or six points of the compass. Not only latitude, but distance from the centre also, should be taken into account. While the centre of a cyclone is yet at a considerable distance and the winds have not yet a great gyratory velocity, they may, even in high latitudes, have a considerable inclination toward the centre, and nearer the equator they may be nearly radial; but even in these latitudes, at places near the centre, where the velocities are very great, the gyrations may become nearly circular.

The preceding is confirmed by rules recently laid down by Dr. Doberck from observations made in the China Sea and in the Philippine Islands. He says:[40]

"The whereabouts of the centre of a typhoon may, in the China Sea, be ascertained by the rule: Stand with your back to the wind, and you will have the centre on your left side, but between 2 and 4 points in front of your left hand. There are, however, certain exceptions to this rule. Thus there often blows a steady easterly gale along the southern coast of China when a typhoon is crossing the China Sea, and the gale blows often steady from northeast about the northern entrance to the Formosa Straits when there is a typhoon in a more southern latitude."

Again he says:[41]

"Further researches have shown that in the Philippine Islands and along the Coast of China as far north as 24° latitude, when you stand

with your back to the wind in a typhoon, you will probably have the centre nearly 4 points in front of your left hand; but on the open sea far from any shore you will generally have it about 3 points in front of your left hand when your ship is in front of the centre of the typhoon, and more than 3 points in front of your left hand behind the centre. Above 25° latitude the angle will probably be found to be between 2 and 3 points. It appears to be smaller the greater the distance from the nearest shore and the greater the latitude. At some distance behind the centre the wind blows generally straight towards it.

182. With regard to the ring of high pressure with its maximum corresponding to the heavy circle in Fig. 1, there are many observations which indicate that it really exists, and is sensible to observation in all well-developed cyclones. This is not shown so clearly from observations made at the same time over different parts of the cyclonic area, and presented in synoptic charts, as from observations made at different times at the same place while the whole system of motions and pressures pass over: though in the former case it is generally observed that around an area of low pressure the barometer stands a little higher than the general average. Before all great storms it is frequently observed that the barometer stands unusually high, and the same soon after it has passed. This was first noticed and remarked upon by Redfield. These are the times when the ring of high pressure around the cyclone in its progressive motion passes over the place of observation. At Havana, Cuba, over which the tropical cyclones of great violence frequently pass, the zone of high pressure is distinctly observed, both before and after the passage of the central part of low pressure and great violence; and the abnormally high pressure which precedes is usually regarded as an indication of the approach of a cyclone. The approach of the hurricane of September, 1875, was indicated at Havana by a sudden rise of the barometer while the cyclone was yet at the Windward Islands, about 1200 miles distant.[41] Also, on the 13th of September, 1876, there was a great rise of the barometer at Havana while a cyclone was causing great destruction on the Island of Porto Rico.[42]

Espy remarks, in his Fourth Meteorological Report, that

"A sudden rise of the barometer, especially in low latitudes, is a proof that there is a storm in the neighborhood, and is one of the first indications of an approaching storm." On the 1st of September, 1845, at Fort Brooke, Florida, at the very time there was a most violent hurricane approaching, and was within 4° or 5° of the place, the barometer arose there that morning 0.15 in., though it had been nearly stationary for more than a month before. A great storm of rain reached there that day; and then in the afternoon the barometer fell 0.3 in. Espy considered it a fact fully established that *without such a rise no storm of dangerous violence need be feared*, but that the converse of it was not true. The barometer rises where the violent part of the storm passes by one side or the other.

We also have evidence of this ring of high barometer from observations in the southern hemisphere. The gale observed at Kerguelen by the exploring party of the *Challenger* "were preceded by an unusually high barometer, which fell rapidly as the storm began from the north; as the wind shifted to the west the barometer rose."

183. The observations of the upper currents of the air are more vague and uncertain. But even here the results of the observations confirm the motions given in the preceding figure for the upper strata of the atmosphere. From 620 observations of the motions of cirrus clouds, Mr. Ley[44] arrived at the following general law showing the relation between the direction of the higher currents of the atmosphere and the distribution of pressure at the earth's surface: *The higher currents of the atmosphere, while moving commonly with the highest pressures, in a general way, on the right of their course, yet manifest a distinct centrifugal tendency over the areas of low pressure, and a centripetal over those of high.*

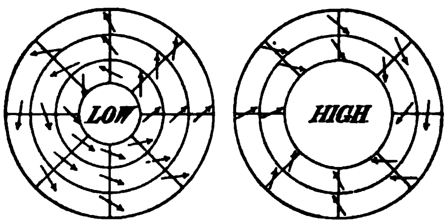
This "centrifugal tendency over the areas of low pressure" is represented by the dotted arrows within the dotted circle of Fig. 1, these observations having been made on the motions of cirrus clouds which are always at high altitudes, where the radial component of motion is outward, but yet generally not so high that the gyrations are anti-cyclonic.

This keen observer of air currents and cloud motions, and generally not seeking a confirmation of any theoretical law, has likewise observed air currents which corroborate in a remarkable manner the counter-motions of the air at the earth's surface and at high altitudes, as delineated in Fig. 1 between the heavy solid and dotted circles. He says:

"There occur at rare intervals in Western Europe depression systems, which affect, but in a very singular way, the directions of the upper currents, reversing them so that they become, on all sides of the area, nearly or quite in opposition to Ballot's law; that is to say, there exists a direct (anti-cyclonic) upper current circulation above a retrograde (cyclonic) circulation of the surface winds."

It is seen from the figure that within a certain belt the currents below and above are reversed. But this is only when the upper currents are at very high altitudes, above the average height of the cirrus clouds. This reversion of the currents is therefore said to occur at only rare intervals; that is, when the cirrus clouds observed are unusually high.

The following figure, given by Hildebrandsson, represents



. Min. Max.

Fig. 4.

the results of his observations of the cirrus clouds at Upsala on the average for the year. The left-hand figure represents the observations for barometric pressures less than 760 mm. and corresponds with the interior part of Fig. 1. The exterior part, those of barometric pressures greater than 760 mm., and corresponds with the exterior part of Fig. 1 beyond the isobar of 760 mm. It is seen that the arrows incline out from low toward high barometer, as in Fig. 1; but in no case do they

indicate completely anti-cyclonic gyrations, as in Mr. Ley's rare cases, the average altitude of the observations not being great enough for that.

From an examination of 121 cases of high winds on Mount Washington, by classifying them with reference to the direction of lowest pressure, Professor Loomis deduced the following conclusions:

"1. High winds on Mount Washington circulate about a low centre as they do near the level of the sea. 2. The motion of the winds is nearly at right angles to the direction of low centre. 3. The low centre at the height of Mount Washington lies behind the low centre at the surface of the earth as much as 200 miles."

At the height only of Mount Washington we would expect to find the winds blowing with reference to the centre of the cyclone very nearly as at sea-level, since it is far below the neutral plane where the radial component of motion is outward from the centre, and especially lower than the altitude at which the gyrations become anti-cyclonic. But as the motions at the neutral plane become circular we would expect them to deviate but little from a circular motion at the top of Mount Washington, and so be nearly at right angles to the direction of the low centre. The third deduction of Professor Loomis is explained by the circumstance that the temperature of the air is not symmetrical on all sides of the cyclone and is colder on the west than the east side of the centre, as represented in § 191, Fig. 5. Taking two points on each side in the line of progressive motion of the centre where the pressures on the earth's surface are equal, as we ascend on the colder side the pressure decreases faster than on the east side on account of the greater density of the air, and so after arriving at a given altitude the pressure on the rear side is less than on the front side. This throws the point of lowest pressure on the level above a little behind that on the earth's surface. On the latter the low centre is perhaps near the medium point between points of equal pressure; but above, it is near the medium point between equal pressures there, and so behind the middle point or point of lowest pressure below.

GRADUAL ENLARGEMENT OF CYCLONES.

184. In the preceding consideration of cyclones it has been supposed that the whole system of circulation has a definite limit, and comprises at all times the same air, which, by means of the vertical circulation, is being continually interchanged between the interior and exterior parts within this limit. This, however, is far from being the case in nature. As long as the vertical circulation is maintained with increasing or at least sufficient energy, the tendency of the cyclone is to extend farther and farther from the centre, and so to continually extend the gyrations and the whole system of circulation over a greater area. But unless the energy which sustains it increases likewise, it must reach a limit, for otherwise the temperature gradient and the forces upon which the vertical circulation depends become barely sufficient to overcome the frictional resistances, when further enlargement must cease; and then, as the energy begins to fail, either through a diminution of the supply of aqueous vapor or a change in the relative temperatures of the lower and upper strata by means of the interchange of air in the vertical circulation, the whole system of circulation must gradually become weaker and finally entirely subside.

It may be that cyclones mostly commence over a small area and gradually enlarge, but this is not necessarily so. If the atmosphere is in an unstable state, and the temperature conditions (§ 156) are such that the initial upward motion takes place over a large area at once, then the vertical and gyratory circulations begin at once over a large area and gradually grow in strength until the frictional resistance from increase of velocity becomes equal to the force, when further increase in violence, as well as extent of limits, must cease. There is therefore a certain limit to the extent and violence of a cyclone somewhat proportional to the amount of energy; and if the initial temperature conditions are such as to start it over a

small area only, it gradually increases in both extent and violence until this limit is reached.

As the area is gradually enlarged and air from greater distances is brought in, there must be such an adjustment between the cyclonic and anti-cyclonic gyrations as to satisfy the principle of § 171, and so the extent of both must increase together, and also that of the ring of highest barometer.

The observed dimensions of the violent part of cyclones are very different, especially in different latitudes. In the Indian Ocean, and especially in the Bay of Bengal, and also in the China Sea, they are said to be comparatively small, ranging from 50 to 250 miles in diameter. According to Redfield the cyclones in the lower latitudes of the western part of the Atlantic Ocean are of about the same dimensions.

Piddington says:[14]

> "In the West Indies the researches of Mr. Redfield and Col. Reid seem to show that though while approaching to, or within, the Islands they are sometimes as small as 100 or 150 miles in diameter, they may, and it would seem most frequently do, after reaching the Atlantic Ocean, dilate considerably, and then often attain to 600 or even 1000 miles in diameter, the wind blowing an excessively severe gale over all this area, and toward its centre becoming of true hurricane violence, and the whole vortex, so whirling, travels over thousands of miles of track."

He further says:

> "The typhoons (cyclones) of the China Sea appear to vary from 60 or 80 miles to three or four degrees in diameter."

In the middle and higher latitudes, however, they are generally from 1000 to 1500 miles, and often much more, in diameter. As a rule the diameters are much greater in the higher than in the lower latitudes. But tropical cyclones are observed at sea and those of high latitudes mostly on land, and those observed at sea have perhaps mostly originated on the continents, and the difference may lie in the places of origin, those originating at sea being generally smaller.

PROGRESSIVE MOTIONS OF CYCLONES.

185. Ordinary cyclones are never stationary, and the directions in which their centres move and their velocities vary not only in different latitudes and regions of the globe, but in the same places at different times. In general their tendency in lower latitudes is westerly, and in the middle and higher latitudes easterly. There are several circumstances which control, to a greater or less extent, the progressive motions of cyclones. The principal one of these is undoubtedly the general motion of the atmosphere in which they exist, not at and near the earth's surface merely, but at high altitudes where the centre of energy is. This carries them along as a stream of water carries along the small whirling eddies which are formed in it. This idea was first suggested by the writer more than a quarter of a century ago.[44] Tropical cyclones move westward, or at least have a large west component, because the general motion of the atmosphere there, up to a considerable altitude, is westerly, while in the higher latitudes cyclones move in an easterly direction with much greater velocities, because there the general motion at all altitudes is easterly, and the velocity, especially at high altitudes, is comparatively great. The average direction for the United States, as determined by Loomis from an examination of the individual directions laid down on the Weather Maps of the Signal Service for three years, comprising 485 cases, is N. 81° E., with an average velocity of 26 miles per hour. The monthly averages indicate an annual inequality. For April to September, inclusive, it is N. 82°.5 E. 23 miles, and for October to March, inclusive, N. 78°.5 E. 29 miles per hour. Hence the direction is more nearly toward the east and the velocity greater in winter than in summer.

The rate of progress of cyclone centres over the North Atlantic Ocean and over Europe seems to be much less, and the annual inequality proportionally less. As the results of later and more extended researches Loomis[17] gives, in miles per hour, the following monthly rates of progressive motion of cyclone centres:

Month.	United States.	Atlantic Ocean, Middle Latitudes.	Europe.
January	33.8	17.4	17.4
February	34.2	19.5	18.0
March	31.5	19.7	17.5
April	27.5	19.4	16.2
May	25.5	16.6	14.7
June	24.4	17.5	15.8
July	24.6	15.8	14.2
August	22.6	16.3	14.0
September	24.7	17.2	17.3
October	27.6	18.7	19.0
November	29.9	20.0	18.6
December	33.4	18.3	17.9
Year	28.4	18.0	16.7

The values for the United States are the averages of 13 years, 1872–1884. Those for the Atlantic Ocean were obtained from the monthly storm tracks published with the International Bulletin for a period of four years, 1879–1882. Those of Europe, from the monthly charts of storm tracks published by the Deutsche Seewarte for five years, 1876–1880. The average in the United States for the 6 winter months is 31.7 miles, while for the summer it is 24.9 miles, the ratio being very nearly as obtained above from the three years.

If the general motions of the atmosphere have a controlling influence upon the progressive motions of cyclones, then we would expect that these motions would not be only westerly in lower latitudes and easterly in the higher latitudes, but also if there is an annual inequality in the one there must also be such in the other. According to the table of § 98 the easterly velocity in the middle latitudes is more than twice as great in January as in July; but the averages for the winter half and the summer half of the year would be nearly as 29 to 23 respectively, the same as the ratio between the winter and summer velocities of the progressive motions of cyclones given above. This velocity, however, is much greater than that of the easterly motion of the atmosphere in the lower strata, even up to a considerable altitude. But the energy of the cyclone, as may be seen from § 159, is mostly above, where the conden-

sation of the aqueous vapor occurs, and where the air temperature in the cyclone differs most from that of the surrounding atmosphere; and so the progressive motion is controlled mostly by that of the general motion of the atmosphere up at that altitude. From the table of § 98 it is seen that at the altitude of 2.5 miles in the United States the east component of the general motion is about 26 miles, the same as the easterly velocity of the cyclones.

In the tropical latitudes, where by the table of § 98 the velocity of the west component of motion of the atmosphere vanishes at no great altitude, especially in the winter season, the westerly velocity of the progressive motion is comparatively small, and must be controlled mostly by the lower part of the atmosphere. It is remarkable that this westerly progression of tropical cyclones occurs almost exclusively at a season when the westerly velocity of the atmosphere extends the highest up.

186. There seems to be a general tendency in cyclones which originate in the lower latitudes to move toward the poles, but observations are wanting to show that this is the case everwhere without exceptions. Such a tendency may arise from the deflecting force due to the earth's rotation being greater for equal gyratory velocities on the polar than on the equatorial sides of cyclones, this force being as the sine of the latitude. Hence the tendency of the gyrating air on the polar side to move toward the pole is greater than that of the air on the equatorial side to move toward the equator, and so the difference between these unbalanced counter forces tends to draw the whole system toward the pole, since the condensation of aqueous vapor and centre of energy are in the cyclonic part, and so the whole system of gyration must follow it. It is evident that such a tendency must, for cyclones of the same extent, be greatest near the equator, where the sines of the latitudes on the two sides differ most, and be comparatively small in high latitudes, where the difference between the sines, even for large differences of latitude, is small.

But the directions and velocities of progressive motion are also much influenced by the irregularities and deflections of

the general motions of the atmosphere, caused by the continents and mountain ranges, as explained in § 123 and following ones. Hence on the east coasts of the United States, of China, and of South Africa, where mostly tropical cyclones pass up into the higher latitudes, they simply drift in the prevailing current, but are probably somewhat aided by the influence referred to above. On the eastern sides of the continents near the tropics, where the deflections and the general currents are mostly toward the equator, as west of Northwest Africa and of Mexico, there is no well authenticated account of cyclones having passed from lower to higher latitudes. We have, however, a reliable account of a very violent and destructive cyclone passing in a direction from the equator toward the pole in the Fiji Islands with a velocity of about ten miles per hour, where there is no known cause for a general current of the atmosphere toward the pole.[48] This occurred on the 3d and 4th of March, 1886. Its track was traced not quite to the parallel of 20° S., but here it seemed to be inclining around toward the east, as if already feeling the influence of the easterly currents of more southerly latitudes. This cyclone was a characteristic one of these latitudes, of small diameter, about 300 miles, minimum pressure of 27.63 in., and of extraordinarily steep gradients, estimated at one inch of mercury in a distance of 63 miles.

187. From the combined action of the two preceding principles, namely, that cyclones have a tendency to move with the general atmospheric current in which they exist, and that they at the same time tend to move toward the poles, all cyclones which originate near the equator must first move in a westerly direction and also toward the pole, after arriving at the parallel of about 30°, which is the dividing line between the westerly and easterly motions of the strata which mostly control the progressive motion of the cyclone, the tendency must be toward the pole only, but after arriving in still higher latitudes, where the general motion of the atmosphere is easterly, the resultant motion must be northeasterly, and finally after reaching still a higher latitude, where the polar tendency is

small, it must be still more toward the east, unless affected by some of the other controlling influences. Hence the path of such a cyclone is somewhat of the form of a parabola, with its vertex at or near the tropical calm-belt, except that the equatorial branch of the parabola continues more on the same parallels of latitude than the polar branch.

Many cyclones of this sort originate during the latter part of summer and the fall in the North Atlantic Ocean, apparently near the equatorial limits of the trade-winds, which at first are comparatively small; but they seem to have great power of continuance, for they usually run through the whole course of their parabolic orbit, gradually expanding as they go, until, after reaching high latitudes in the northern part of the Atlantic, they have the usual dimensions of cyclones of these latitudes. Being carried at first in a nearly west direction by the westerly motion of the atmosphere in the trade-wind zone, until they arrive in the neighborhood of the West India Islands and the Gulf of Mexico, they then curve around toward the north over the United States, or along the east coast and the Gulf Stream, up into higher latitudes, aided here in their polar movement, no doubt, by the deflected current referred to in § 123. Here their progressive motions are controlled mostly by the strong easterly tendency of the atmosphere in these latitudes, and so their directions become more toward the east, the same as those of cyclones having their origins in the higher latitudes.

The rate of westerly progress of these cyclones is comparatively small while near the equator, and it becomes still smaller while they are curving around into the higher latitudes; but after arriving into the middle and higher latitudes, where their general direction is a little north of east, they have the usual velocities of progression of cyclones generally of these latitudes. According to Loomis[17] the average direction and velocity of forty of these cyclones while moving westerly was respectively 26° north of west and 17.4 miles per hour. In only two cases was the direction south of west. Of course the farther west, the more the directions inclined away from the equator. The

average direction of these storms while moving eastwardly to the parallel of 40° was E. 38°.5 N., and the hourly velocity in this part of their course was 20.5 miles. In the first part of this, however, the course was more northerly and gradually changed around to a course only a little north of east.

The steady current of the northeast trade-winds is not favorable to the origination of cyclones, and therefore they seem to originate at the northern limit of the calm-belt only, or at least not so far within that the deflecting force depending upon the earth's rotation is too small to produce gyrations; for, we have seen, this force becomes very small near and vanishes at the equator. Hence, of the forty paths of cyclone centres examined by Loomis, no part of any one of them was found within 10° of the equator. If the conditions of a vertical circulation were found at the equator, they would not give rise to any cyclone, and there would be little violence and no sensible barometric depression in the centre. In all seasons of the year except during the latter part of summer and the fall, the trade-winds blow down to a latitude where the deflecting force of the earth's rotation is too weak to give rise to cyclones, and as they cannot originate within the zone of these winds, they can generally originate at the time only in which these winds do not extend down very near the equator.

188. The east coast of China, and the adjacent oceans, including Japan, is also traversed by numerous cyclones of this sort, originating in the same manner and mostly at the same season of the year. These likewise curve around in a kind of parabolic path with its vertex near the parallel of 30°, until they arrive in the higher latitudes and assume a direction nearly east across the North Pacific Ocean. The rate of progress is probably about the same as in the case of the tropical cyclones of the North Atlantic Ocean and the United States.

In the Arabian Sea and Bay of Bengal there are similar small cyclones originating near the equator, which first move only a little north of west, and then gradually curve around toward the north and are lost within the continent. But these seem to be interfered with by the great Asiatic monsoon, so

that they originate mostly spring and fall at the times of the changes of the monsoon, when there is a month or more of comparatively calm weather. Consequently there is a tendency toward two maxima in the monthly numbers of cyclone frequency, as is seen in the following table. The average westerly velocity of these cyclones, as obtained by Loomis[17], is only 8.5 miles per hour.

On the eastern coast of Africa and the adjacent ocean, including Madagascar, cyclones originating in the Indian Ocean near the southern limit of the equatorial calm-belt, when it has its most southern position in midwinter of the northern hemisphere, being then forced down to the parallel of about 12° S. (§ 142), pass around into middle latitudes of the southern hemisphere, where their directions become more nearly east, being controlled there by the strong general easterly currents of the atmosphere in these latitudes. The forms of their paths are very similar to those of the North Atlantic and North Pacific Oceans, but their motions in them are comparatively very slow, usually only 3 or 4 miles per hour. These being in the opposite hemisphere, the time most favorable for their origination, and consequently the time of greatest frequency, is in February intead of August.

The relative frequency of the occurrence of all the tropical cyclones in the several months of the year may be seen in the table on page 282.

In the South Atlantic Ocean the southeast trade-winds often extend considerably beyond the equator, and at any season too near the equator for cyclones of this class to originate. Hence on the east side of South America and on the adjacent ocean there do not seem to be any cyclones which move around in a parabolic orbit into the higher latitudes of the southern hemisphere.

189. Although cyclones are carried along by the general drift of the atmosphere, yet as the velocity of this is very different at different altitudes, varying circumstances of the cyclone may cause great variations in the progressive velocity, while that of the general drift remains the same. According

THE YEARLY PERIODS OF CYCLONE FREQUENCY IN SEVERAL SEAS.[49]

	Arabian Sea.	Bay of Bengal.	S. Indian Ocean.	Java Sea.	China Sea.	Havana.
No. of years.	234	139	40	..	85	363
No. of cyclones.	70	115	53	12	214	355
Authority.	Chambers.	Blanford.	Piddington, Thom and Reid.	Piddington and Thom.	Schück.	Poey.
Jan.	6	2	17	25	2	1
Feb	4	0	25	42	0	2
Mar.	3	2	19	8	2	3
April.	13	8	15	8	2	3
May.	18	16	7	0	5	1
June.	29	9	0	0	5	3
July.	3	3	0	0	10	12
Aug.	3	4	0	0	19	27
Sept.	4	5	2	0	27	23
Oct.	6	27	2	0	16	17
Nov	14	16	7	0	8	5
Dec.	3	8	6	17	3	2

as the central controlling power of the cyclone is at greater or less altitudes must the progressive velocity be greater or less, so far as it depends upon the mere drift of atmosphere; and these variations depend very much upon the hygrometric state, as well as the unstability, of the atmosphere, as may be seen from the table of § 159. Taking the first example in which the vertical temperature gradient is represented by the column A, it is seen by comparing the column C in the case of saturation with that of A, and then that of C′ in the case of $\tau - d = 4°$, that the weight of the differences, and consequently the power of the cyclone, is much lower in the former case than in the latter. The more nearly, therefore, the air is saturated, other circumstances remaining the same, the lower down is the controlling power, and consequently the less must be the progressive easterly velocity of cyclones in the middle latitudes.

The annual inequalities in the velocities of the progressive motions of cyclones, we have seen, is less than that of the general easterly velocities of the wind; and this may be caused by the atmosphere's being more nearly saturated in winter than in summer. It may also be the cause of the greater velocity of progressive motion over the United States at all

seasons than over the Atlantic Ocean and Europe, since the climate of the United States is continental, and consequently drier, than that of the Atlantic Ocean and of Europe, which, for reasons given in § 122, has somewhat of an oceanic climate. Over the United States, therefore, the vapor has to ascend a little higher generally before condensation takes place, and so the seat of controlling energy is higher, and the progressive velocities of cyclones greater, than on the Atlantic Ocean and in Europe.

But the more unstable the state of the atmosphere is, the higher is the central controlling power, as may be seen from the same table of § 159; and so the rate of progression also depends upon this circumstance, and consequently it causes variations at different seasons and places.

190. As the energy of the cyclone is mostly in the aqueous vapor condensed, and without this we rarely have the conditions of more than an initial cyclonic action, the velocity and direction of the progressive motion of a cyclone depends, to some extent at least, upon the distribution of this vapor in the region in which the cyclone exists: for the cyclone is likely to be drawn somewhat in the direction in which there is the most vapor, and to pass around those regions in which there is little vapor. It is for this reason, perhaps, that the chain of lakes between Canada and the United States seems to be a great highway for cyclones. It is also very often observed that cyclones in the interior of the United States, especially in winter, instead of pursuing their usual easterly direction, strike directly across in a southeasterly direction to the Atlantic coast where the air is warmer and the aqueous vapor more abundant, being drawn in the direction from which they receive the most sustenance, and then proceed along the coast toward Newfoundland. In the winter season, also, cyclones which make their first appearance on the Pacific coast of the United States, if not too far north, instead of passing through the cold and dry interior of the continent, follow the coast down into Lower California and Mexico, and, crossing over into Texas and the gulf region, then make their way up along the eastern coast of

the United States, as is usual with all tropical cyclones of that region coming from the direction of the West India Islands, thus seeming to skirt entirely around the interior of the continent on the warmer and moister side of the continent, because here more aqueous vapor is found for their support.

We have seen that there are two dry belts extending entirely around the globe (§ 120), except so far as they are broken up by monsoons and the deflected atmospheric currents of continents and mountain ranges (§§ 124, 125). It is doubtful whether the polar tendency of tropical cyclones is in general strong enough to carry them through the trade-wind zones against the trade-winds, and whether the amount of vapor in the dry belts is sufficient to sustain them in their passage through them into higher latitudes, except on the eastern sides of the continents, where the deflected atmospheric currents counteract and change the directions of the trade winds and carry aqueous vapor with them for the support of the cyclone. Hence the regions of these deflected currents on the east sides of the United States and China, and of the southern part of Africa, seem to be gaps in the dry belts through which alone, or at least for the most part, tropical cyclones are able to pass into the higher latitudes. We have had one notable example, however (§ 186), in which such a passage seems to have been made, but this was a cyclone of unusual power and violence; and the passage was through a region where there are numerous and very large islands.

191. In all which precedes upon this subject, as well as upon the subject of cyclones in general, no account has been taken of the effect of differences of temperature of the atmosphere upon the polar and equatorial sides of the cyclone, but the whole region of atmosphere in which the cyclone exists is supposed to be of uniform temperature, except within the cyclonic region. But in the real cases of nature, especially in the middle latitudes and in the winter season, there is a great difference of temperature, where the cyclone is large, between the polar and equatorial sides, which has nothing to do with the temperature disturbance upon which the origination of the

cyclone depends, but which give rise to large modifying influences after the cyclone has originated, which would not take place in an atmosphere of uniform temperature on all sides around the cyclone.

The following figure represents the action of a large cyclone, say 1000 miles in diameter, in the middle latitudes, in winter, at which time the difference of temperature on the two sides may be as much as 20° C. On the southeasterly and easterly sides the cyclonic currents, as represented by the arrows, carry the warmer and moister air of lower latitudes to higher ones,

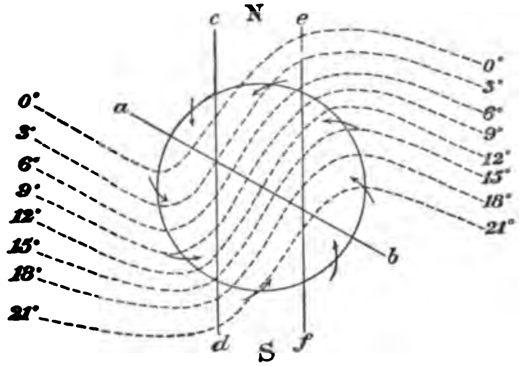


Fig. 5.

and on the northwesterly and westerly sides, the colder and drier air of higher into lower latitudes. The difference of temperatures between the west and east sides of the cyclones is also increased by terrestrial radiation through the clear air on the west side, which is the clearing-up side of the cyclone, while on the other, clouds prevail which hinder this radiation. The effect from both causes upon the isotherms, which we may at first suppose to have extended nearly in an east and west direction, is somewhat as shown by the figure. The temperature gradient now is in the direction of the line *a b* instead of from the pole toward the equator, and is steeper, the isotherms being closer. Comparing now the temperatures on the same latitudes in the lines *c d* and *e f*, as indicated by the disturbed isotherms, it is seen there is a great difference and

a steep temperature gradient in an east and west direction, where, before cyclonic disturbance, there was none. Not only does the cold and dry air on the west side tend to press the warmer and moister air on the east side still farther toward the east, but the latter is continually forming a new centre of temperature disturbance a little in advance, which becomes a new cyclone centre, and so there is an apparent progressive motion from the tendency to continually form a new cyclone a little in advance of the old one. Such an effect would evidently take place in an atmosphere without any drifting progressive motion; but this effect must be subordinate to, and much less than, that of the progressive motion of the atmosphere; for, if it were not, since its action is always in a direction from west to east, cyclones in the tropical latitudes would never move westwardly.

The effect of a cyclone upon the isotherms, as represented in the preceding figure, may be seen almost any time upon the synoptic weather charts of the Signal Service, in cases of well-developed cyclones; but of course the observed effects are never so regular as those of the preceding figure, since we never have a perfectly regular cyclone, as assumed above, and besides there are always other abnormal disturbances of temperature combined with those of the cyclone.

The air on the east side of a cyclone, at least near the surface of the earth, is very damp and oppressive, much more so than it was while in the lower latitudes from which it has been drawn; for as it is carried up to higher latitudes over a surface which becomes colder with increase of latitude, it of course cools down some, though not nearly to the normal of the latitude where the air remains undisturbed, and in this cooling it becomes more nearly or quite saturated. Exactly the reverse of this takes place on the west side. The air found there is not only colder because it has been brought down from a higher latitude, but it is much drier than it was in its original position.

VEERING AND BACKING OF THE WIND AND CHANGES OF PRESSURE AND TEMPERATURE.

192. If a cyclone remained stationary without any progressive motion, there would be no veering of the wind, and the only change would be a gradual increase and then decrease in its velocity. The barometric depression, and the temperature of the interior, mostly in the upper strata, would be subject to similar changes, and these changes would have a period coinciding with the time of duration of the cyclone. But where the cyclone has a progressive motion these changes follow in more rapid succession, and complete their period in the time that the cyclone occupies in passing over the place of observation.

Let the arrows of the following figure represent the direc-

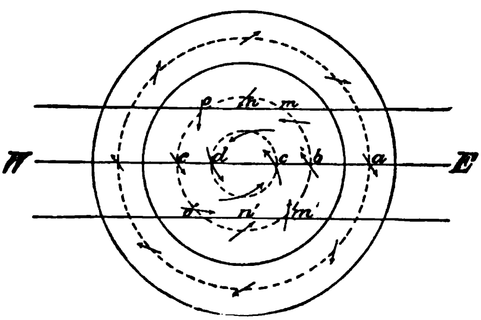


Fig. 6.

tions and relative velocities of the wind in a cyclone, the interior solid circle representing the line of highest barometric pressure between the cyclonic and the anti-cyclonic gyrations. If the cyclone passes centrally over the place, say from west to east, then while the place is in the outer or anti-cyclonic part, as at *a*, there is a very slight rise of the barometer and a very gentle wind from a point west of north, if there are no abnormal disturbances. Then comes the highest barometric pressure and the calm connected with it, and when the point *b* of the cyclone has reached the place of observation the direction of

the wind is southeasterly, and the velocity continues to increase with little change of direction, and the barometer to fall, until the place of observation relative to the cyclone is at some point *c* still nearer the centre, where the cyclonic motion is most rapid. While the centre, and contiguous parts very near on either side, are passing over the place of observation, there is a calm, called the "dead calm," and the barometric pressure is then the least. When the progressive motion has continued a little longer, until the place of observation is at *d*, there has been a complete change of the wind to the contrary direction, and the barometer is now rising, which continues until the opposite side of the ring of high pressure and of calm arrives. If the direction of progressive motion is exactly from west to east, the change in the direction of the wind at the centre is from about S.S.E. to N.N.W., but if, as most usual in the middle latitudes of the northern hemisphere, the cyclone moves toward the E.N.E., then the change, it is readily seen from the figure, must be very nearly from S.E. to N.W. At the time of this sudden change of direction near the centre there is of course a great and sudden change in the character of the air. On the east side, for reasons given in § 191, it is warm, damp, and oppressive, and this is suddenly changed to cold and very dry air, using the terms warm and cold here relatively to the seasons.

Within the tropics, where the direction of progressive motion in the northern hemisphere is W.N.W., the manner of veering is readily seen by placing the figure with the line W.E. in that direction. The directions of the arrows then indicate that the change, when the centre passes over the place of observation, is from a northerly to a southerly direction. The same may be done for any other directions of progressive motion. When the place of observation, in any case, is at *e*, the violence of the wind has very much diminished and the height of the barometer increased, and soon after follows the highest barometer with its accompanying calm, after which the usual light variable winds, due to various abnormal disturbances, are observed, the normal inversion of direction again in the anti-cyclonic

part being usually entirely obscured by these. If the area of central calm is small and the gyrations are very rapid close up to the centre, as in violent cyclones of moderate dimensions at sea, then there is a very sudden change of a very strong wind to the opposite direction.

If the *north side* of the cyclone passes eastwardly over any place of observation in the northern hemisphere, so that the place of observation occupies successively the places within the cyclone *m*, *n*, and *o*, then it is seen that the change of the more violent winds in the cyclone proper is from about E. at *m* to N.E. at *n* and then to N. at *o*. In this case there is a change of the wind through only about one quadrant, and this in a direction contrary to the motion of the hands of a watch, called *backing;* but if the south side of the cyclone passes in the same direction, over the place of observation, so that this occupies successively the places *m'*, *n'*, and *o'* in the cyclone, then the change is from S. at *m'* to S.W. at *n'*, and then to W. at *o'*, and the change in this case, through about one quadrant, being in the direction of the hands of a watch, is called *veering*.

The direction in which the wind changes depends entirely upon which side of the place of observation the centre of the cyclone passes; and the amount of change, upon the distance of the path of this centre from the place of observation. If the distance is small in comparison with the dimensions of the cyclone, the change is through about two quadrants, but with greater distances the amount of change is less. For other directions of progressive motion, the veering or backing of the wind is indicated in all cases by the arrows, when the line W.E. is placed so as to coincide with these directions.

The winds at any given place are liable to change as frequently, on the average, in one direction as the other, unless the cyclones pass more frequently on the one side of the place of observation than the other. In the middle latitudes of the United States and of Europe cyclones pass more frequently on the north side, so that in these latitudes the change is more frequently in the same manner as the motion of the hands of a watch. It was only for this reason Dove found a preponder-

ance of changes of this kind, in confirmation of his theory with regard to the rotation of the winds.

For the southern hemisphere it is only necessary, in what precedes, to change N. to S. and *vice versa*, in order to have the successive changes in the veering and backing of the wind in each case.

193. In what precedes with regard to the changing of the directions of the winds, no account is taken of the effect of the secondary conditions introduced by the action of the cyclone itself, where there is a considerable difference of the general temperature on the two sides, by which a steep temperature and pressure gradient is introduced in an east and west direction between the lines *c d* and *e f*, as represented in Fig. 5. The other component of the gradient in a north and south direction remains much as it was at first before there was any cyclonic disturbance, and has nothing to do with the cyclonic disturbance, but simply with the general easterly motion of the atmosphere, and has already been considered in a previous chapter. But the other component of the temperature gradient introduced has to be here considered in connection with the cyclone. The tendency of the cold dry air on the west side is to press eastward below even beyond the centre of the cyclone, and as cold and warm portions of air do not readily intermingle, but tend to keep apart, there is often a long line, several hundred miles in length, in the central, or perhaps rather easterly, part of the cyclone, where there is a short but sharp temperature and pressure gradient, and where, in the passage of a cyclone eastward, there is a very sudden change of the wind around to some westerly or northwesterly point, and there may even be violent and simultaneous squalls, called *line squalls*, along this whole line with a very sudden change of temperature. And the general effect of this east and west temperature gradient on the west side of the cyclone, beyond the line of sudden transition from warm to cold air, is to give rise to a corresponding pressure gradient, and so to cause the wind to be from a direction more nearly west than it would be by a purely cyclonic motion. It is on account of this increased pressure

gradient on the west side, no doubt, that Loomis obtained a greater inclination and velocity of the wind in the west than in the east quadrant of a cyclone. His results fcr the several quadrants were (*Sill. Jour.*, July, 1874):

	W. Quadrant.	S. Quadrant.	E. Quadrant.	N. Quadrant.
Inclination......	58° 48′	49° 35′	32° 6′	47° 27′
Velocity in miles per hour......	10.1	8.8	8.3	7.6

The line of sudden transition from warm to cold air, and likewise of the meeting of the southeasterly and northwesterly winds, is a line of minimum pressure, and the low-pressure area on both sides is sometimes called a "trough." It is simply a great extension of the central area of low pressure and of calm of a regular cyclone, so that the winds, instead of coming from all sides in toward and around the centre, come in, not directly, but obliquely, toward the middle of the trough. For this reason the isobars of cyclones in the middle latitudes are not circular but somewhat elliptical, with the northerly part of their longer axes corresponding to the trough, extending in a direction a little to the east of north, as was first discovered by Espy, and now more fully proven by the researches of Loomis." He found from an actual measurement of the greatest and least diameters of the isobars represented on the Signal Service maps during a period of three years, that the average ratio of the longest diameter of the isobars to the shortest was nearly as two to one.

Since the preceding secondary effect depends upon differences of general temperature in latitude, they occur principally and most strikingly in the middle latitudes and in the winter season, but they may occur at all latitudes and in all seasons wherever conditions are such, whether of temperature or terrestrial radiation, as to cause a difference of temperature between the two sides; for it must be remembered that slight differences of temperature and small pressure gradients give rise to violent winds where there are no deflecting forces arising from the earth's rotation to counteract the forces of these

gradients, as in the case of the interchanging motions in the cases of the vertical circulations of the atmosphere and of cyclones. In the lower latitudes, however, where the difference of temperature depending upon latitude is small, these effects are very small, and hence it is said that in the tropical cyclones there is an "absence of any marked squall or change of weather during the passage of the trough in the tropics, that is, at the moment when the barometer begins to turn upward.[50]"

In the easterly progression of cyclones in the middle latitudes, following one another in somewhat regular succession at intervals of a few days, there is first encountered the southerly winds on the easterly side of the trough, and then, in a few days, the northerly winds on the west side, and this is frequently repeated at short intervals. This seeming struggle, as regarded by Dove, between "equatorial and polar winds," now so well understood, was formerly a great mystery to the meteorologist.

CYCLONE OF AUGUST 2, 1837, AT ST. THOMAS.

194. As an example of some of the theoretical results arrived at, we give (page 293) the observations of a cyclone at St. Thomas, made by Hoskiaer, and copied here from Dove.[51]

The normal barometric pressure here being about 29.9 inches, the whole low-pressure area, and the part of the cyclone having any considerable violence, passed over the place of observation in about 24 hours. This, at the average rate of 15 miles per hour, at which cyclones progress here, would make the diameter of this part 360 miles, and consequently, according to Piddington (§ 184), it was not one of the smallest kind. With a barometric depression of nearly two inches in the centre, even with this area, the barometric gradients become very large. Between 6 h. 30 m. and 7 h. 30 m. the change of barometric pressure was 0.977 in., which, with the assumed rate of progressive motion, would make a gradient of nearly one inch in 15 miles. On the opposite or rear side at about the same distance from the centre, the gradient is less, being only about

Mean Time.			Barometer.	Direction of the Wind.	
	h.	m.	Inches.		
Aug. 1	18	0	29.932		
Aug. 2	2	10	.754	N.W.	Gale freshening.
	3	20	.666	N.	
	3	45	.666	N.	
	4	45	.489	N.	
	5	40	.443	N.E.	
	5	45	.310	N.E.	Hurricane.
	6	30	29.133	N.W.	
	6	35	28.911	N.W.	
	6	45	.778	N.W.	
	7	0	.778	N.W.	
	7	10	.600	N.W.	
	7	22	.289	N.W.	
	7	30	.156	N.W.	
	7	35	.111		Dead calm.
	7	52	.067		
	8	10	.067		
	8	20	.067		
	8	23	.423	S.S.E.	Hurricane.
	8	33	.511	S.E.	
	8	38	.600	S.E.	
	8	45	.689	S.E.	
	8	50	.778	S.E.	
	9	0	28.995	S.E.	
	9	10	29.133	S.E.	
	9	25	.222	S.E.	
	9	35	.310	S.E.	
	9	50	.399	S.E.	
	10	10	.489	S.E.	
	10	35	.577	S.E.	
	11	10	.599	S.E.	
	11	30	.621	S.E.	
	14	45	.754	S.E.	
	20	0	.888	S.W.	
	21	0	29.910	E.	

0.8 in. in 15 miles. These gradients, therefore, are enormously greater than the estimated gradient of the cyclone of the Fiji Islands (§ 186), although the minimum of the depression in this latter case was much lower. But it is seen that the barometer, as usual in such storms, was irregular, being affected by various irregular and abnormal disturbances, so that the gradients at any given time can be only approximately determined. In the front part of the storm at 7 h., taking the average between 6 h. 30 m. and 7 h. 30 m., it was approximately one inch in 15 miles or 21 mm. in the distance of 60 geographical

miles. Assuming the exact centre of the cyclone to be at the place of observation at 8 h. 10 m. and that the progressive velocity was 15 miles per hour, the distance of that gradient from the centre was about 17 miles, or 27 kilometers. Putting now G_4, § 57, equal to 21 mm. and $r = 27{,}000$ meters, and $l = 18°$, the value of v which satisfies the expression for a temperature of 30° C. is 24 meters per second, or about 54 miles per hour for the approximate gyratory component of velocity, and as the inclination in such small and violent cyclones is small, even in this latitude, this is not much less than the real velocity. Of course this estimate is subject to all the uncertainties of gradient and distance, where there are so many irregularities in the barometer, and where the exact progressive velocity is not known.

By comparing the value of $\nu = v/r$ in the preceding computation, with that of $2n \sin l$ as deduced from Table V, it is seen that the former is more than 14 times greater, and hence the gradient here in this low latitude depends almost entirely upon the centrifugal force of the gyrations, and very little upon the earth's rotation.

As the progressive motions of cyclones in this vicinity are in a direction only a little north of west, and this cyclone seems to have passed very nearly centrally over the place of observation, by placing the line W.E. in Fig. 6 in that direction, and making allowance for a greater inclining in toward the centre in this low latitude than that represented in the figure, which is more adapted to the middle latitudes, it is seen that in the more violent part of the front of the cyclone, as at the points e and d, the wind is N.W., as represented in the preceding table of observations. Then comes the "dead calm" at the centre with lowest pressure, which continues about 45 minutes, and so, with a progressive velocity of 15 miles per hour, it must have been about 12 miles in diameter; but of course there is no definite limit, since the gyrations at first become merely perceptible, and then gradually increase with distance from the centre, though very rapidly, so that at only a short distance they become extremely violent. At the points c and b in the rear

part there is a complete change of direction to the S.E., as shown in the preceding observations, which has taken place in a short time, while the central calm passed over the place.

The observations, as given, were not commenced soon enough, and continued long enough, to indicate the belt of high pressure and the anti-cyclonic directions of the wind on the outer border of the cyclonic region of disturbance, but these, most probably, would have been masked by small abnormal disturbances.

MANILLA TYPHOON OF NOVEMBER 5, 1882.

195. This cyclone has been investigated by Loomis,[17] and the following table of observations of its central and more violent part is copied approximately from the curves of his diagram.

Time.		Bar. Pressure.	Temperature.	Wind.	
				Velocity.	Direction.
d.	h.	mm.		m.	
4	22	752	23°	6	N.W.
	23	52	23	10	N.W.
5	0	51	22	20	W. ½ N.W.
	1	50	22	21	W.N.W.
	2	48	22	30	W. ½ N.W.
	3	47	21	33	W.N.W.
	4	46	21	22	W.N.W.
	5	46	21	34	W.N.W.
	6	46	21	28	W.N.W.
	7	45	21	29	W.N.W.
	8	43	21	36	W.N.W.
	9	37	21	44	W.N.W.
	10	35	22	2	N.N.E.
	11	37	25	9	E. ½ S.E.
	12	38	24	12	S-E. ½ E.
	13	43	26	25	S.E. ½ E.
	14	45	27	28	E.S.E.
	15	46	26	30	E. ½ S.E.
	16	47	25	22	E.S.E.
	17	48	25	16	E.S.E.
	18	49	25	15	E.S.E.
	19	50	25	14	E.S.E.
	20	51	25	10	E.S.E.
	21	52	24	10	E.
	22	52	24	12	E.S.E.

The minimum pressure of this cyclone was not very low nor the gradients very steep, except very near the centre, although the velocities of the wind were considerable. This is because the cyclone was large for this latitude, and the distances mostly were much greater than in the case of the St. Thomas cyclone. It is seen from the expression of G_4, § 57, that with larger values of r the value of G becomes smaller. The sudden reversion of the direction of the wind is explained as in the preceding case. This cyclone, however, being a little nearer the equator, the directions of the wind seem to have been more nearly radial, as they should be, than in the preceding case, so that the change is from W.N.W. to E.S.E. mostly, instead of from N.W. to S.E., as in the case of the St. Thomas cyclone. The duration of the central calm seems, as usual, to have been short.

If we assume that 5 d. 10 h. was the exact time of the passage of the centre, and compare the several pressures and temperatures for each hour preceding and following this passage, we get the following results:

Hours Before and After.	Pressure.		Temperature.		Wind Velocity.	
	Before.	After.	Before.	After.	Before.	After.
	mm.	mm.				
1	737	737	21°	25°	44	9
2	43	38	21	24	36	12
3	45	43	21	26	29	25
4	46	45	21	27	28	28
5	46	46	21	26	34	30
6	46	47	21	25	22	22
7	47	48	21	25	33	16
8	48	49	22	25	30	15
9	50	50	22	25	21	14
10	51	51	22	25	20	10
11	52	52	23	24	10	10
12	52	52	23	24	6	12

From this it is seen that the rear side of the cyclone is the warmer side, and in the passage of the cyclone the transition is from cold to warm, contrary to what occurs in the middle latitudes, as illustrated by Fig. 5 and shown by observation; and the isotherms, if drawn, would curve down on the front side

toward the equator instead of toward the pole. This is because the progressive motion is nearly in the contrary direction. The effect, however, is comparatively small, since in this low latitude, even in November, the difference of temperature due to difference of latitude on the two sides is small. The corresponding increase of pressure on the front side on account of lower temperature seems to be only very faintly indicated in the preceding table. In general the wind velocities in the preceding table are greater on the front than the rear side, but this is most probably due to the winter monsoon, which at this season is already set in, and the direction in this region is rather from the N.W., which tends to increase the velocity in front a little and to decrease it a little in the rear.

CYCLONE OF CIENFUEGOS ON SEPT. 5, 1882.

196. The following observations of this cyclone are taken from the Report of the Chief Signal Officer for 1883, p. 759.

Time.		Barometer.	Temperature.	Wind.	Force.	State of the Weather.
d.	h.	Inches.				
5	0	29.82	82°	N.	2	Clouds between 1° and 2° quadrant.
2		29.76	80	N.W.	3	Heavy gusts of rain.
8		.66	80	N.W.	3	Do.
9		.59	78	N.W.	4	Hurricanes, gusts.
9	30′	.50	78	W.N.W.	4	Violent gusts of wind and rain.
10	0	.38	78	W.N.W.	4	Hurricane.
10	15	.31	78	W. ¼ N.W.	4	Gloomy and cloudy weather.
10	30	.27	78	W.	4	
10	45	.22	78	W.	4	Greatest intensity.
11	0	.18	78	W. ¼ S.W.	4	Blowing with great force.
11	30	.13	78	W.S.W.	4	Do.
11	45	.15	78	S.W. ¼ W.	4	Barometer rising.
12	0	.16	78	S.S.W.	4	Gloomy appearance, rain in torrents.
12	30	.18	78	S.S.W.	4	Do.
1	0	.27	78	S.S.W.	4	Do.
1	30	.32	78	S. ¼ S.W.	4	Do.
2	0	.44	78	S. ¼ S.W.	4	Less wind gusts and of short duration.
2	30	.46	78	S. ¼ S.W.	4	
3	0	.46	78	S.	4	Gusts diminishing.
3	30	.49	78	S.	4	Rapid gusts.
4	0	.55	78	S.S.E.	4	Heavy rain.
5	0	.60	78	S S.E.	4	Do.
6	0	.66	78	S.E.	4	Rains and gusts.
7	0	.70	78	S.E.	4	Strong winds with lasting gusts.
8	0	.78	78	S.E.	4	Do.
9	0	.79	78	S.E.	4	Heavy rain and strong wind.
10	0	.80	78	S.E.	3	Gusts at intervals.

This is an example in which the centre of the cyclone did not pass over the place of observation, but on the north side.

There is, accordingly, no dead calm and sudden change of the wind to the opposite direction, but a gradual backing of the wind from a N.W. wind in the front to a S.E. wind in the rear. The explanation of this is readily seen by reference to the accompanying figure, in which the large arrow indicates the direction of progression, and the parallel line with the small arrows and the even hours of the time indicates the place of observation at the several times with reference to the centre of

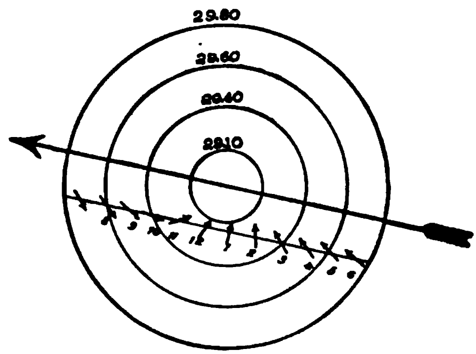


Fig. 7.

the cyclone. The calm centre passed north of the place of observation, and so when it was nearest, the place of observation was at the distance of greatest violence. If the centre had passed on the other side there would have been a gradual veering instead of backing of the wind.

RAIN AND CLOUD AREAS IN CYCLONES.

197. Since the air flows in from all sides of a cyclone toward the central area, and consequently gradually ascends there, this must be a region of cloud and rain, just as the equatorial belt is, where the air, coming in from both sides in the lower strata, is deflected upward and becomes an ascending current. The explanation of the formation of cloud and rain, and the height to which the surface vapor has to ascend before cloud-formation takes place, is the same in both cases, and has been already given (§ 111).

The rain and cloud areas do not coincide with, nor are they

even concentric with, the areas of low pressure and of ascending currents; for the ascending and partially condensed vapor and the cloud may be, and generally is in the middle and higher latitudes, carried eastward beyond the limits of the low pressure area, by the strong easterly currents above, into the anticyclonic area before it falls as rain; and hence the centre of the area of cloud and rain falls considerably in advance of the centre of low pressure and of cyclonic gyration in their easterly progression. These areas, as those of low pressure, are elliptical, and not circular as they would be in case of regular conditions with no greater progressive motion of the atmosphere above than below.

According to Loomis,[11] "the form of these rain areas is sometimes quite irregular, but generally it approximates to an ellipse of which the major axis is not quite double the minor axis." The average distance of the centre of rainfall from the centre of low pressure, north of latitude 36°, was found to be 300 miles, but it sometimes exceeded 750 miles. The general direction of the longer axis of the rain-area, and of the centre of this area from that of the area of low pressure, is the same as the direction of the progressive motion of cyclones. This indicates that both the former and the latter depend mostly upon the general motion of the atmosphere. The whole cyclone and storm area drifts with the average velocity of the different strata, but the upper strata, in which the vapor condensation takes place mostly, have a greater average velocity, and so the rain and the cloud are carried forward, and the rain falls mostly in the front part of the cyclone in the direction in which the general atmosphere and the whole cyclone are moving. Of course this is only approximately so, for we have seen (§ 189) that the direction of the progressive motion may be affected considerably by other circumstances. The direction and distance of the centre of rainfall from the centre of low area indicates the location of greatest energy in the cyclone, and this, we have seen, does not depend entirely upon the mere drifting of the upper strata of the atmosphere where the centre of energy is.

In general the cloud area, though somewhat concentric with that of the rain area, is very much larger; for it requires a considerable depth of cloud to give rise to rainfall, and so the outer parts of the cloud, arising mostly from the lateral expansion of the air in its ascent, and not directly from the ascent of air, are not of sufficient depth to furnish rain. In the easterly progression of a cyclone, therefore, there is observed first merely haziness, then a thin and light stratum of cloud, and then the heavy, deep nimbus, from which rain falls.

Where the conditions of a cyclone are such that the vapor of the air is carried by the ascending current in the interior to very high altitudes, where there is a freezing temperature, the fine cirrus clouds formed by the freezing of the vapor there are carried eastward far in advance of the cyclone, by the swift-moving easterly currents of these altitudes in the form of fine filaments, which are often so distorted by the small whirling motions which frequently exist in the atmosphere as to resemble a horse's tail, and are called "mares' tails." These cirrus clouds of various forms, coming from a westerly direction, while the hazy border of the cloud has not yet made its appearance, are generally precursors of a violent cyclone and storm coming from that quarter, which may be expected in a few days.

198. It has been shown (§ 156) that the conditions of a cyclone are not absolutely dependent upon the condensation of aqueous vapor. There may, therefore, be considerable cyclonic action and barometric depression, with little or no rainfall. If the atmosphere were entirely void of vapor, but was in the unstable state for dry air, from some slight determining cause arising from a small local increase of temperature over a considerable area, moderate vertical and cyclonic circulations would ensue, and likewise barometrical depression in the central part. Also where the temperature over a considerable area is considerably above that of the surrounding air, as shown in § 159, it gives rise to cyclonic action of some degree of strength, even where the vapor is not sufficient to produce clouds of sufficient depth and density to give rise to rainfall.

The energy of the cyclone is in the condensation and not in the rainfall, the latter being only an evidence of the former, but there may be considerable cloud-formation without rainfall.

In 101 cases of low barometer in which the depth of rain did not amount to one eighth of an inch, Loomis[61] found that more than one half showed a barometric pressure less than 29.70 inches; more than one third were below 29.60 inches, and nearly one fourth of the cases were below 29.50 inches. He says:

"There seems to be no room for doubt that barometric minima sometimes form with very little rain, and continue without any considerable rain for eight hours, and sometimes for twenty-four hours and longer. These barometric minima seldom continue stationary for eight hours, but almost invariably travel to the eastward."

In general, areas of low barometer are areas of cloud and rain but there is no proportionate relation between the depth of rain, and of barometric depression, for the latter depends very much upon the deflecting force of the earth's rotation, and the extent of area, being the result of an integration of the gradient through a distance from the centre, while the former depends mostly upon the velocity of the ascending currents and the amount of moisture in the air at any given place, and not upon extent of area. Where considerable depressions are observed with little or no rainfall, the cyclonic action and the barometric gradients are generally small, but the depression is considerable from the extent of area.

199. There is a characteristic difference between tropical cyclones and those of the middle and higher latitudes. Abercromby says:[66]

"Cirrus and halo appear all round a tropical cyclone, while they are never seen in the rear of a European storm; and though the way in which the rain seems to grow out of the air in front of the cyclone is the same everywhere, the sky and clouds in the rear of a hurricane are much softer and dirtier than in temperate cyclones. There is not that sharp difference between the quality of clouds in front and rear which is so striking in higher latitudes. Still greater is the absence of any marked squall or change of weather during the passage of the trough in the tropics,—that is, at the moment when the barometer begins to turn

upwards. Some who study hurricanes have scarcely noticed any change then ; and all are agreed that the trough-phenomena are slight."

According to Padre Viñes[11] the approach of cyclones from the E. or S.E. is always indicated at Havana by the appearance of cirrus clouds while the vortex is 500 miles or more distant, and while fair weather is yet prevailing. The same is also observed as they pass off at a distance. This indicates that the cirrus clouds are observed on all sides of the tropical cyclones, in accordance with what is stated above. Again, according to Dr. Doberck,[11] at Hong Kong and in the China Sea "there does not generally exist a fine-weather area behind the cyclone, as the S. and particularly the S.W. winds blow there very fresh, accompanied by overcast, damp and frequently wet weather. Thunder-storms likewise follow after a typhoon, especially along the coast of Southern China."

In the middle latitudes the general motion of the air is easterly at all altitudes, but its velocity is much greater above than below. Hence cirrus clouds and halos never appear in the rear of cyclones in Europe, or anywhere in the middle latitudes. Near the equator the general motion of the atmosphere is westerly in the lower part, and at the season of the year in which the tropical cyclones mostly appear, it is so up to a high altitude, and above that altitude the velocity of easterly motion is comparatively small. In this latitude, therefore, the air ascends more nearly vertically, and in the expansion of the air in all directions as it comes under less pressure, the vapor of the ascending current is diffused somewhat equally in all directions around the centre of the cyclone, so that there is here no striking difference between the clouds in front and rear, and at Havana the cirrus clouds are observed when the cyclone centre is still 500 miles or more distant toward the east, and while fair weather is still prevailing.

The reason the usual trough-phenomena observed in higher latitudes in the passage of a cyclone over a place are not observed in tropical cyclones is clear from what has been stated with regard to the Manilla typhoon (§ 195). In these there is no close contact between warm and cold air of very different tem-

peratures and a sudden change from the former to the latter, but, on the contrary, the change is from colder air to that which is only a little warmer.

In the higher latitudes the west side is the clearing-up side of the cyclone, because the strong easterly currents above, especially at high altitudes, carry all the vapor and cloud eastward, so that soon after the centre of the cyclone has passed a place, the clouds mostly have passed it also, and clear weather prevails. It is not so in tropical cyclones. Here the progressive motion is westerly, and although the atmospheric currents below are also westerly up to a considerable altitude, yet the westerly velocity decreases with increase of altitude, and becomes easterly above, so that the clouds are not driven ahead in advance of the cyclone, but there is enough of easterly current above in the cloud region to carry the clouds eastward to the rear of the cyclone, and consequently there is there "overcast, damp and frequently wet, weather," instead of a fine-weather area such as is observed in the rear of cyclones in higher latitudes.

RESULTANTS OF CYCLONIC AND PROGRESSIVE MOTIONS.

200. Where the atmosphere in which a cyclone exists has a progressive motion in some direction, as it generally has, the resultant of this and the cyclonic motion differs from the purely cyclonic motion, and the effect of the progressive motion upon the velocity and direction of the cyclonic motion is different on different sides of the cyclone.

Let *ab*, Fig. 8, represent the velocity and direction of the cyclonic motion on the several sides of the cyclone, with its centre progressing a little north of east, as usual in the middle latitudes, and as represented by the arrow, and also let *bc* represent the velocity and direction of the wind at the earth's surface, which may be assumed to be the same, or nearly, in all parts of the cyclone, and to have very nearly the same direction as that of the cyclone centre, but its velocity is generally much less. It is seen from the figure that the resultants, represented

by *ac*, indicate very different velocities and inclinations on the different sides of the cyclone. On the side of the cyclone on the right-hand side of the centre's path the velocity is changed from *ab* to *ac*, and is, therefore, considerably increased, though the direction is changed but little, since the directions of both the cyclonic and progressive motions are nearly the same. On the left hand, on the contrary, both the velocity and direction, as represented by *ac* as compared with *ab*, are very much

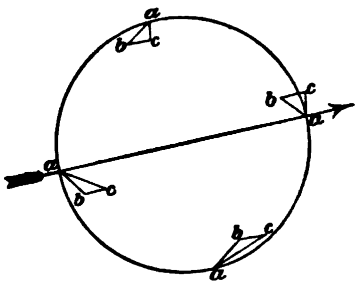


Fig. 8.

changed, the former being decreased and the latter becoming nearly radial in a direction toward the centre. On the front side the direction only is very much changed, since nearly the whole effect of the progressive motion of the air is upon the direction of motion and not upon the velocity. The inclination here, with the assumed relation between progressive and cyclonic velocity, is negative or outward. In the rear the effect is an increase of both velocity of the wind and its inclination toward the centre.

Of course the relations between the resultant velocities and directions and those of the cyclonic, depend upon those between the two components. If the progressive velocity of the air is small in comparison with the cyclonic, as is usually the case on land, the cyclonic velocities and directions are not much changed, but this is generally different at sea, where the progressive velocity of the air is usually greater. The relations, as represented here in Fig. 8, are perhaps those which usually occur at sea, where the progressive velocity may be nearly as great as the cyclonic.

201. The progressive motion of the air in the region of the West India Islands being comparatively large, its effect upon both the velocities and directions is quite large. Here the trade-winds become nearly east winds (§ 123). Hence, the winds whose cyclonic component of motion is from the north or south suffer the greatest deviation from the original cyclonic direction, the inclination toward the centre being decreased where the wind is from the north and increased where it is from the south. This will be better understood from Fig. 9, in which *ab* represents the cyclonic and *bc* the progressive

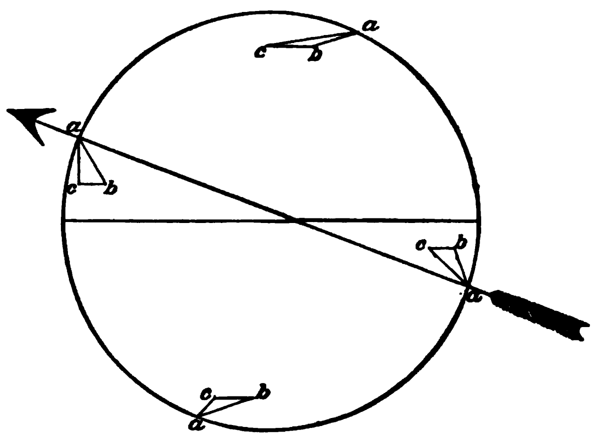


Fig. 9.

component of the motion of the air at the earth's surface, the inclination here in these low latitudes being assumed to be about 45°. The direction of progressive motion of the centres of the cyclones here is about W.N.W., as represented by the arrow in the figure. It is seen that in front of the storm the inclination of the resultant *ac* is diminished and becomes nearly at right angles to the radius, while in the rear it is increased and becomes nearly radial, in consequence of the progressive motion of the air *bc*. On the right-hand side the direction of the progressive motion coincides nearly with that of the cyclonic motion, and hence the direction is little changed, but the velocity is greatly increased; but on the left-hand side

the direction and velocity are both much changed, the latter being very small, since the cyclonic motion is somewhat counteracted by the progressive motion.

These results are verified by observations of the hurricanes of the Antilles of 1875 and 1876. In the hurricane of September, 1875, it is said,[11] "the winds of the anterior part of the storm were approximately circular, or with a slight inclination toward the centre in some cases." But from observations made at numerous places "the winds of the second (S.E.) quadrant, which remained at all these places when the vortex was at a considerable distance, suffered a great deviation toward the centre, and in some cases, likewise, the winds not far from the vortex." This large deviation is represented by *ac* in the rear part of the storm in the figure. The reason that this great inclination occurred in *some cases* only near the vortex, is that there the cyclonic component is large and has but little inclination, the winds there, as we have seen (§ 175), becoming nearly circular. Again, on the island of Porto Rico "little deviation in the winds of the third and fourth (S.W. and N.W.) quadrants was noted, some greater convergency in those of the first (N.E.) quadrant, and a great inclination toward the centre in those of the E. and S., especially as they became at a great distance from the vortex."

Of the hurricane of the 19th of October, 1876, it is likewise said,[12]

"After its passage by Havana the winds, which in the posterior part blew from west to south, suffered a great deviation toward the centre, and that not only at a distance from the vortex, but even in its vicinity."

And—

"The winds which prevailed from S.S.E. to S., and which, during the passage of the vortex by Havana blew with force in the different towns situated to the E.S.E. of Havana, suffered likewise a very notable inclination toward the centre. With respect to the winds which prevailed in the first quadrant in the different localities of the anterior part of the cyclone, there was observed in them likewise a notable convergency, though in general in a less degree."

It is readily seen from a mere inspection of Fig. 9, why on

the Island of Porto Rico the winds had little deviation from the tangent in the S.W. and N.W. quadrants, that is, in the anterior parts of the cyclone, and why the inclinations in those of the east and south quadrants were great, especially at a great distance from the centre, where the inclination of the cyclonic component is large (§ 175).

Also why the winds in the posterior of the cyclone of the 19th of October, after it had passed Havana, suffered a great deviation toward the centre, and the winds in the region E.S.E. of Havana, which prevailed from E.S.E. to S., suffered likewise a very notable inclination.

202. From what precedes it is seen that the navigator, in determining the direction of the centre of a cyclone from the direction of the wind, should, in addition to considering latitude, distance from centre, and velocity (§ 181), likewise consider in what quadrant of the cyclone he is situated, since the direction of the centre with reference to that of the wind is so different in different quadrants, especially where there is a large progressive motion of the air, as in the trade-wind regions. In the front part of the cyclone, where the mariner is in the greatest danger, the direction of the centre is generally more nearly at right angles to the direction of the wind, and consequently the old rules of the circular theory are more nearly correct here than on any other side of the storm; while in the rear the motions are more nearly radial in the direction of the centre. This difference between the front and rear of the cyclone is embraced in Dr. Doberck's rule for the Philippine Islands (§ 181).

It is seen from Figs. 8 and 9 that both in the middle and tropical latitudes the velocity of cyclonic motion is increased on the right-hand side of the path of the centre, and decreased on the left-hand side, by the progressive motion of the air in the neighborhood of the cyclone. The right-hand side, therefore, in the northern hemisphere has long been recognized as the *dangerous side* of the cyclone. In the southern hemisphere it is of course reversed. In endeavoring, therefore, to escape the most dangerous part of the cyclone, care should be taken to avoid if possible this side.

203. But the effects of the progressive motions of the air at a considerable altitude above sea-level, on mountain tops and in the regions of the clouds, especially the cirrus clouds, are much greater than at the surface, because the velocities of progressive motion are much greater. In fact up at the altitude of the cirrus clouds the progressive component is usually the principal one, and the cyclonic merely causes considerable perturbations in the strong easterly current of these regions. This is seen from the results deduced from the observations of the cirrus clouds made at Zi-ka-wei, as given in § 83. From these it is seen that the motions are mostly from a westerly direction, the cyclonic component being merely sufficient, for the most part, to cause deviations of one or two points from this direction, and rarely sufficient to entirely reverse this strong easterly current and give rise to a motion from an easterly direction. And this is in accordance with the estimates made by Mr. Ley from his early observations on the motions of the cirrus clouds. He states that the most elevated ones not uncommonly traverse a distance of 120 miles in an hour. This is, no doubt, when the great easterly motions of the air in these regions coincide in direction with a great cyclonic motion, as in the S. or S.E. quadrant. On the other hand, he states that calms are uncommon in this elevated stratum, and that he observed only twice an actually motionless cirrus cloud." In these cases the observations, no doubt, were made in the N. and N.W. octant of a cyclone region, where the cyclonic and progressive motions very nearly or quite counteract each other.

204. Professor Loomis' discussion of observations of the Signal Service, made on the top of Mount Washington at the times of low barometer, gives the following velocities and inclination of the wind :

	West Quadrant.	South Quadrant.	East Quadrant.	North Quadrant.
Velocity in miles.......	49	44	37	32
Inclination...........	55° 7′	− 13° 25′	− 53° 44′	69° 54′

Fig. 10 is a copy of his graphic representation of the directions of the resultants. Comparing this with Fig. 8, which may be regarded as a representation of the resultant motions usually at the surface of the ocean, it is seen that the results are somewhat similar, but the effect of the progressive motions upon the directions of the cyclonic are more striking, the inclination in the front part of the cyclone being outward or negative. But the resultant velocity deduced from these results is from N. 65° W. 28 miles per hour, while in Fig. 8 the direction of progressive motion was assumed to be from a point a little S. of W. If we assume that the cyclonic velocity on the average was 30 miles per hour and the inclination at this alti-

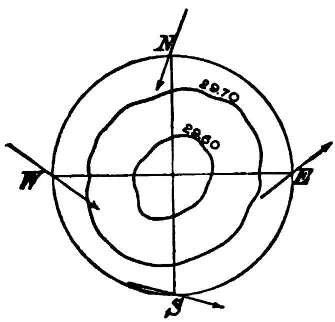


Fig. 10.

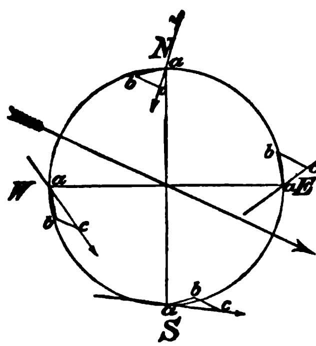


Fig. 11.

tude 10°, the results of such an assumption with the progressive velocity and the direction given above are represented by Fig. 11. By comparing this with the preceding figure, it is seen that the results are very nearly the same, and so are accounted for by a cyclonic motion as assumed above.

In cyclonic motions at a considerable altitude the inclination is outward and not inward, as represented in Fig. 8; and so for the same velocities of progressive motion the inclination of the resultant *ac* in front is greater and in the rear less above than at the earth's surface.

205. We come now to the examination of the results obtained by Mr. Ley from observations of the upper currents already given in the table of § 180. From a mere inspection of

the angles given in this table for the several districts it is readily seen that they result from a cyclonic motion with a considerable outward inclination, combined with a large progressive motion in the direction of the progress of the area of depression, or nearly so, represented by the large arrow in Fig. 3. Let *A*, *B*, *C*, etc., in Fig. 12 represent the centres of the several octants in Fig. 3, and let *Aa*, *Ba*, *Ca*, etc., represent the velocities of cyclonic motion, having an outward declination of about 30°; and also *ab* represent the velocity of progressive motion of the upper atmosphere in the direction of the motion of the centre of low-pressure area, rep-

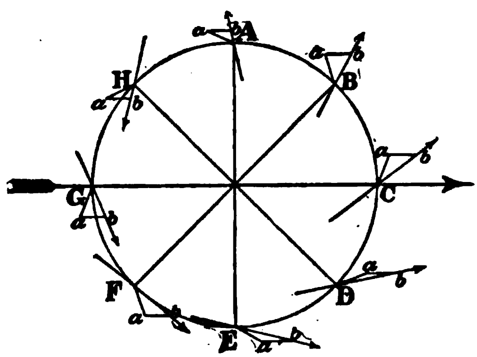


Fig. 12.

resented by the arrow, and let us also suppose that this latter velocity is very nearly equal to that of the cyclonic velocity *Aa*, *Ba*, *Ca*, etc. The resultants will then be represented by *Ab*, *Bb*, *Cb*, etc. It is readily seen that the effect of the progressive component is to increase the angles with the radius in the front and to diminish them in the rear. The effect is similar to that on Mt. Washington represented in Fig. 10. It is seen that the largest angles with radius are in the districts *B*, *C*, and *D*, and the smallest those of *F*, *G*, and *H*. By a reference to the column of "Mean angles with radius" in the table of § 180, it is seen that this is very nearly the case, there being two exceptions in the inner districts, in which the maximum in front and the minimum in the rear seem to be a little more toward the north and south sides respectively. In

vague observations of this sort it cannot, of course, be expected that any hypothesis would give a nice agreement with observation in quantitative results.

It is seen from Fig. 12 that the resultant velocity in the octant *A* is very small, and consequently the average angle from observation very uncertain. In fact, the progressive velocity so nearly counteracts the cyclonic, that this must be a district, on the average, of calms rather than of currents. This is in accordance with the observations of Mr. Ley, who with regard to this district remarks that "the upper current, which had previously nearly coincided in direction with the trajectory, presently changes, as a rule, to a nearly opposite point, having just before become very slow. In the interval the cirrus, when visible, which is rarely the case, is sometimes stationary, sometimes moves towards, and sometimes from, the centre of depression, these three instances being nearly equally common, but calms and motions toward the centre predominating." [40]

THE EYE OF THE STORM.

206. In all parts of the world there appears to be in the central part of most at least, if not all, cyclones a thinning of the cloud-stratum and a partial, and sometimes complete, clearing away of the clouds. This was observed by Dove at Königsberg as early as the year 1827.[18] He says:

"Whenever the equatorial current sets in suddenly and is as suddenly displaced again, at the precise time when the barometer is at its lowest level, the showers which belong to the equatorial current are separated from those produced by the intrusion of the polar current into it by a short period of clear weather, to which I gave the name of 'clear interval.' At the centre of a cyclone, when the barometer is lowest, a similar fine moment is so frequently observed that sailors have given it a special name, 'the eye of the storm.' Captain Salis describes its occurrence in a cyclone experienced by the ship *Paquebot des Mers du Sud*, in latitude 38° S., longitude 22° E., in these words: 'While there was a dense bank of clouds all round us, the sky over our heads was quite clear, so that we could see the stars, one of which, right over the head of the foremast, was so bright that every one on board noticed it. The barometer was 27.79 inches English.'"

Dove here has in mind the old idea of a constant wrestling between equatorial and polar winds, and makes a distinction between what he observed and what is seen in the central part of tropical cyclones. But in fact he simply observed what occurred in the passage of the centre of cyclones, modified however in high latitudes by the trough-phenomena (§ 193).

The following quotation taken from an account of a cyclone in the Arabian Sea, by Dr. Malcolmson, is given by Piddington:

"During the height of the storm the rain fell in torrents, the lightning darted in awful vividness from the intensely dark masses of clouds that pressed down, as it were, on the troubled sea. *In the zenith there was visible an obscure circle of imperfect light of ten or twelve degrees.*"

With regard to the cyclone of October, 1848, in the Bay of Bengal Piddington further says:

"The superintendent of the light-house at False Point, Palmiras, distinctly states that at the time of the passage of the centre, or for about two hours of calm, the stars were seen very clear overhead, with a thick bank of haze all round."

This feature of cyclones seems to belong mostly to tropical latitudes. Abercromby[60] says:

"The tropical cyclone has a striking feature which is absent in our latitudes: There is a patch of blue sky over the calm centre, which is well known in hurricane countries as the 'eye of the storm,' or as a 'bull's-eye.'"

From the fact that it was noticed by Dove, the phenomenon does not seem to be entirely absent in higher latitudes, though it is undoubtedly less distinctly manifested than in tropical latitudes; and the name is not so appropriate as it is in the case of the small tropical cyclones in which the clear patch seems to be small and often very distinct, and which consequently gave origin to the name.

Every one perhaps has observed in ordinary rain-storms that after it has rained for some time there is frequently a cessation of rain at least, if not a partial clearing away of clouds, as though the storm had all passed by; but after a short time

there is a thickening and a darkening of the clouds and a second shower of rain. Indeed, of so frequent occurrence is it that it is generally remarked that the latter is the "clearing-up shower," as though it were a matter of course, and to be expected before the final clearing away.

207. The origin and explanation of these phenomena are to be found in temperature conditions, such as assumed in the table of § 159, which give rise to a vertical circulation which does not extend to the top of the atmosphere, often to a moderate height only, and which, except so far as it acts upon the upper part of the air by friction merely, gives rise to cyclonic motion in the lower strata only. It was the opinion of Redfield that our great revolving storms do not generally extend to a greater altitude than a mile, and this is no doubt sometimes the case; for the cirrus clouds sometimes seen through an opening of the lower clouds, the eye of the storm, frequently appear to be undisturbed by the storm beneath. But at other times, we know,—and this is frequently the case,—the highest cirrus clouds are brought into the whirl, and have a cyclonic motion. If the conditions are such in the upper part of the atmosphere that the ascending current becomes colder instead of warmer than the surrounding part of the atmosphere, of course it ceases to ascend and is deflected off horizontally in all directions before reaching that level. As the gyratory motion depends upon the vertical circulation, this, at least at first, extends up to this level only, but the strata above may be acted upon more or less by friction. The air, charged with moisture, in its vertical circulation, in toward the centre below, up in the interior to a given altitude, and outward above in the middle strata, necessarily moves in a path somewhat elliptical; so that it is being deflected outward above and still ascends until at a considerable distance from the centre; and so there is little condensation of vapor in the central part, and the cloud stratum is thin, sometimes entirely wanting. And this state is still further promoted by the gyratory motion, which is confined mostly to the lower and middle strata, bringing the air down from above, it may be, down pretty low, into the inte-

rior central part, where it is carried out horizontally on all sides; and the descending air in the interior above is of course clear air. The effect under such conditions is a thinning of the cloud and a scantiness of condensation and rainfall in the centre, a ring of deeper and denser cloud at some distance from the centre, which gradually shades off to the outer limit. Where, however, the conditions give rise to a vertical or cyclonic circulation up to very high altitudes, this phenomenon is perhaps never observed.

The tendency of the air in the interior of a cyclone to settle down, and thus to either partially hinder the ascending currents and the cloud-formation, and so to cause a thinning of the cloud, or to even cause a gently descending current and a complete clearing off here, is no doubt caused often by the falling of a great amount of comparatively cold rain from the upper strata, which, both from its weight, which is partially sustained by the air in falling, and also from the cooling effect, increases the downward pressure in the centre. And this would especially be the case in tropical cyclones, where the air ascends more vertically and symmetrically on all sides, and the rain is not mostly carried away in front of the centre before it falls. Hence the central partially or wholly clear space is mostly observed in these cyclones.

In the regular progression of a cyclone in the middle latitudes somewhat centrally over a place, the cloud and rain area of the front part, extending far toward the east, first passes over, occupying a half-day, or a day and more, and then the front part of the ring of dense cloud with a heavy shower of rainfall. After this there are indications of a clearing up, and even the sun may break through the cloud for an hour or two: but presently there is an apparent gathering and thickening of the cloud and a second shower. This is at the time of the passage of the rear side of the ring of denser cloud. After this there is the final clearing up.

It is not to be supposed that this is a regular occurrence in every rain-storm, or that the cloud ring in the special cases of favorable conditions is regularly formed on all sides, but simply

that in general the tendency is toward the formation of such a ring.

SECONDARY CYCLONES.

208. It often happens that the conditions of a smaller cyclone, such as described in § 157, or even several of them, are contained within the limits of a larger one. The isobars then, which in a perfectly regular cyclone are circular, become very irregular. The contained cyclone may be such as to give a secondary minimum pressure, or it may simply produce derangements in the regularity of the isobars of the primary cyclone. Where the isobars become crowded, and consequently the barometric gradients steeper, the velocities are likewise greater; for in such places the motions of the primary and secondary cyclones are somewhat in the same direction. Where, however, the isobars are farther apart, and the gradients consequently smaller, the velocities are also smaller, since there the motions are nearly in contrary directions, and partially or quite neutralize each other.

Fig. 13 represents the effects of secondary cyclones contained in a primary, upon the isobars and the directions of the wind. In the one on the left the depression is not sufficient to give rise to a secondary minimum of pressure, but merely to cause considerable derangement of the isobars, crowding them together on the one side and widening them on the other, and changing somewhat the velocities and directions of the wind. The one on the right being more violent and the depression deeper, produces a secondary centre of low pressure, with the wind blowing around it, but with much greater velocity on the right side, where the directions of the currents of both the larger and the smaller cyclone coincide, than on the other side where the cyclonic motion of the smaller cyclone is but little greater than that of the larger in the contrary direction. The ring of high barometer of the large cyclone is increased in height in an irregular manner by those of the smaller ones falling upon it. There are usually many smaller secondaries of this sort, which tend to distort the isobars, and to cause irregu-

larities in the velocities and directions of the wind as the cyclone and its secondaries pass over any place of observation.

Secondary cyclones occur mostly in the southern and eastern quadrants of a cyclone, in which the air is warmer and moister than in the other quadrants, and where consequently the conditions are more favorable for the origination and maintenance of these smaller cyclones. They are rarely, if ever, found in the other quadrants.

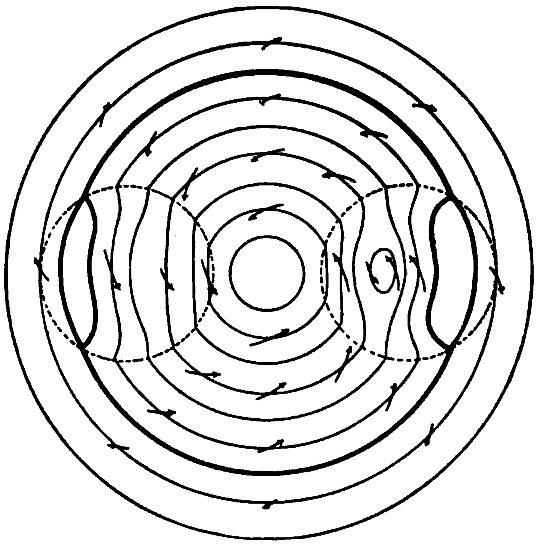


Fig. 13.

209. Of course each secondary cyclone gives rise to an area of more dense cloud and more abundant rainfall within the large one. The amount of rainfall, therefore, in a large cyclone may be very unequally distributed, large amounts of rain falling in some places, while at others at no great distance off, little or none falls. In fact, the larger cyclone may be such as to give rise to cloudiness mostly, and to little or no rain, while there are numerous small secondaries contained within it, which cause heavy rainfalls over limited areas. In the progressive motion of a primary cyclone, with its numerous small secondaries, each one of the latter, as it arrives at any place,

causes an increase in the density and darkness of the clouds, with a shower of rain of longer or shorter duration, and greater or less abundance, after which the general cloudiness, or perhaps very moderate rainfall, prevails until a new secondary arrives, when the same thing occurs again. Such weather is called *showery weather*. After the whole of the large cyclone has passed over, the showers cease and the weather for a while becomes settled.

While all areas of low barometer, whether of primary or secondary cyclones, are generally areas of more or less rainfall, or at least of cloudy weather, it is not to be supposed that there may not be much cloudy weather and rain without any sensible barometric depression. The atmosphere over a large area of the earth's surface may be nearly saturated with aqueous vapor, and nearly or quite in a state of unstable equilibrium, and yet the temperature conditions may not be such as to give rise to a vertical and a cyclonic circulation over an area sufficiently large to cause a sensible depression in the middle, and yet the vertical circulation and the ascension of moist air may be sufficient to give rise to much rain. For it must be remembered that a barometric depression depends, not only upon gyratory circulation, but also upon extent of area, and where the area is small, the barometric depression becomes sensible only in the case of rapid gyrations, to which the conditions may, or may not, give rise. When the atmosphere is in the condition described above there are often many local rains, with little cyclonic action, and consequently little wind, or damp, sultry weather may prevail with little or no rain. But if, after a time, the temperature conditions become such as to cause an initial inflowing of air over a large area toward some central point, and a vertical and cyclonic circulation over this area, this gives rise to a cyclone of considerable extent, and when this, by its usual progressive motion, has passed away in a day or two, the damp drizzly weather is changed and followed by fair weather.

STATIONARY CYCLONES.

210. If, for some reason, there is a local and permanent cause of temperature disturbance between the central and exterior part of any somewhat circular portion of the atmosphere, we have the conditions of a cyclone which is independent of the state of unstable equilibrium, but which would be aided and strengthened by such a state. The cause of the temperature disturbance being fixed, of course the cyclone cannot have a progressive motion.

The rough approximate conditions of such a cyclone are found in the northern part of the North Atlantic Ocean, where there is a large area over which the temperature is higher than that of the surrounding parts, and this is especially the case in the winter season. The central part of this area of higher temperature is near Iceland, where the mean annual temperature is about 16° F. above the normals of latitude for the year, given in the table of § 68. In January, however, the temperature on the continents in these latitudes has become very much less, while that of the ocean has changed but little, so that at that time the abnormals of temperature are at least twice as great, and likewise the difference between the temperature of this area and that of the surrounding parts of the atmosphere, so that now we have the conditions of a cyclone of about twice as much energy as in the case of the average for the year, or during the spring and fall. But in July the reverse takes place; there is then an equalizing somewhat of the temperatures in all longitudes, and the abnormals of temperature are very small.

Similar conditions are found in the northern part of the North Pacific Ocean, but not so marked. The greatest abnormal temperature for the mean of the year is only about 8° F., and that for January a little more than twice as much. Hence in the northern parts of both the Atlantic and Pacific Oceans we have roughly the conditions of a permanent cyclone at all seasons except the summer season, the effects of which upon

the winds and the barometric pressures of these regions are observed, especially during the winter season.

In the North Atlantic Ocean the cyclonic winds on the southern side of this great and permanent cyclone in winter, with its centre near Iceland, coinciding in direction with the general westerly winds of these latitudes on the Atlantic Ocean, produce unusually strong westerly winds at this season; but adjacent to Great Britain and the coast of Norway they come more from a southwesterly direction, and then, curving around toward the extreme northern part of the ocean, and on the northern side of the cyclone, in Greenland and adjacent to it on the ocean, the winds are from the N.E. This cyclonic motion causes a barometric depression in the centre in winter of about 10 mm. below the normal mean pressure of the latitude.

In the summer season the temperature of that region is so nearly the same as that of the surrounding regions, and of the normal temperature of latitude, that there is no sensible cyclonic action or barometric depression, and the winds which prevail there are the westerly winds of the general motions of the atmosphere in these latitudes.

211. In the region of the cirrus clouds above the winter cyclone of the lower strata of the atmosphere over the North Atlantic Ocean, the motion of the air, in the exterior part at least, is anti-cyclonic, and consequently over the British Isles and Europe generally the winds have a considerable north component, which, combined with the large east component of the general motion of the atmosphere at high altitudes, causes their directions to be from a point more to the N. of W.; while in the summer season, when there is little or no cyclonic effect, the general westerly winds at high altitudes are not sensibly disturbed. There is, therefore, an annual inequality in the observed directions of the cirrus clouds over these regions, the directions being, according to theory, from a point a little more to the N. of W. in winter than in summer. This is found to be the case from observation. Taking the averages

of the angles ψ for each month, given by the late J. A. Brown, § 180, we get:

$$\text{October—April,} \quad \psi = 284^\circ$$
$$\text{May—September,} \quad \psi = 268^\circ.$$

Hence it is seen that during the summer half of the year the currents are almost exactly from the west, as they should be, for if entirely undisturbed by the anti-cyclonic action they should be from a point a little S. of W., since the motion of the atmosphere above is slightly poleward. But during the winter half year they are from a point 16° more toward the north, because now there is a northerly component arising from the anti-cyclone of the cirrus regions.

A similar result has been obtained from observations made on the upper clouds at many places in Europe, by Dr. Hugo Hildebrand Hildebrandsson.[60] His results are contained in the following table:

Zones.	Winter.	Summer.	Year.
Below 760 mm. (minimum)	W. 7° 42′ N.	W. 11° 30′ S.	W. 2° 24′ S.
Above 760 mm. (maximum)	W. 42 12 N.	W. 23 24 N.	W. 31 0 N.
Mean................	W. 16 54 N.	W. 3 24 N.	W. 10 54 N.

Taking the general mean, we find that the direction in winter is from a westerly point 13° 30′ more toward the north in winter than in summer, and that in the latter season it is very nearly from the west, as it should be. Taking the results for the two zones separately, it is seen that in each the difference between winter and summer is very nearly the same, but that for the average of the whole year the direction is from a westerly point 33° 24′ farther north in the pressures above 760 mm. than in those below 760 mm. This is in accordance with theory; for the low pressures are more in the interior of the great cyclone, where the gyrations above, at least in the lower cloud-region, may not only have little anti-cyclonic motion, but even some cyclonic motion, and hence the directions should be from a point farther south than in the outer border, in the zone of

high pressures, where the great anti-cyclone above may extend over all Europe.

Of course, the individual observations are affected by the varying pressures of every transient cyclone passing along, but in the averages of great numbers of observations the effects of these are mostly eliminated, and we have left mostly that of the great stationary cyclone only.

212. The conditions of an ordinary stationary cyclone are found in summer on every island in the ocean; for since the annual range of temperature on land is much greater than that of the ocean, the mean annual temperature on land and sea being very nearly the same, the island in summer, especially in high latitudes where the annual temperature changes are great, becomes very much warmer than the surrounding ocean. And this is not merely an incipient condition, such as is required generally to originate a cyclone when the other conditions of proper vertical temperature gradient and hygrometric state are present, but it is continuous, especially during the day, and keeps up the vertical and cyclonic circulations independently of the other conditions. Of course this condition is confined to the island, and so there is no progressive motion. The cyclonic action, however, is of no great violence, for the temperature differences and gradients are mostly in the lower strata of the atmosphere, and in the great body of air in the upper strata they are small. Their power, therefore, to give rise to a vertical, and so to a cyclonic, circulation is small, but this is very much increased where the interior of the island consists of highlands, as has been explained in § 132 in the case of monsoons.

Australia furnishes such conditions in some measure, but the island comprises too great a range of latitude for a perfect cyclone, which requires that the area shall be of smaller extent; and besides it is too near the equator for the influence of the earth's rotation to be considerable, especially on its equatorial side. The effect, therefore, arising from the temperature conditions here has been treated as monsoonic and not cyclonic.

though some allowance has been made for the deflecting force of the earth's rotation upon the directions of the winds.

During the summer season the great continents of Europe, Asia and America are heated up considerably above the surrounding regions, which give rise to vertical circulations, inward below and outward above, giving rise to the summer monsoons, and causing considerable barometric depressions in the interior, especially in Asia, but the areas are of too great extent and comprise too great a range of latitude for much cyclonic effect to be produced, and the lower pressures in the interior are to be attributed almost entirely to the differences of temperature, and very little, if any, to cyclonic gyrations. The influence, however, of the deflecting force of the earth's rotation upon the directions of the monsoon winds is no doubt considerable.

COLD WAVES AND NORTHERS.

218. It frequently happens in the United States during the winter season that there are large and progressive areas of intense cold which originate mostly in the extreme Northwest and progress easterly and southeasterly, frequently as far as the Atlantic Ocean. On account of the great contrast between the temperature of this air and that which it encounters, its easterly progression causes very great and sudden changes of temperature, and hence it is called a "cold wave." Those down in Texas are generally called "Northers."

These cold waves have been investigated by Lieut. Woodruff[67] of the Signal Service, for the years 1881–84. He finds that they occur in winter only, and more in January than in any other month; that 86 per cent originate east of the Rocky Mountains and 14 per cent come across these mountains, and that by far the greatest number travel across the country from Helena, Montana, to the Atlantic in 32 to 40 hours. He gives instances of temperature changes of 40° to 50° F. in 24 hours.

These cold waves are closely connected with, and form a part of, the cyclones which pass over the United States, and an explanation of them has been already partially given in the

explanation of the trough-phenomena and weather sequences in the passage of a cyclone over any given place (§ 193). The whole cold area, however, is not due entirely to the whirling of colder air from higher to lower latitudes on the west side of the cyclone, but also, in a great measure, to the increased terrestrial radiation in winter through the clear air on the clearing-up side of the cyclone, especially on the high plateaus east of the Rocky Mountains. The easterly progressive motion of this cold air west of the Mississippi River is increased by the easterly slope of this plateau, upon the same principles that winter monsoons and land winds are, in the case of high and sloping plateaus in the interior of a country; so that the cold stratum is actually forced under a cyclone a little beyond the centre, but cannot advance faster than the cyclone which it follows, and upon which it in a great measure depends for its origination and continuance.

214. That a cold wave is simply the rear part of a cyclone with the degree of cold more than usually intensified by certain favorable circumstances, is evident from the fact that we have preceding and accompanying it the same sequences of temperature, wind, and pressure which are observed in high latitudes in the passage of cyclones in the winter season, when the trough-phenomena are developed. There are first southeasterly and southerly winds, moist air, increasing cloudiness and rain, and low pressure; then follow a sudden change of the wind to the north or northwest, a rapid increase of pressure, and a lowering of the temperature. Lieut. Woodruff says:[57]—

> "In studies upon the origin and movements of areas of high and low barometer it has been shown generally that they move almost invariably across the United States from west to east. The determination of the movement of the area of low barometer largely determines the movement of the following high. Now most of the areas of low barometer are formed in the region east of the Rocky Mountains, and as these areas move eastwardly the high moves in, and we have accompanying a cold wave of more or less intensity. Even if the low area pursue an abnormal track the circulation of the winds about the high and low is such as to produce almost invariably a decided fall in temperature, if the low be eastward of the high."

In the general easterly progression of the atmosphere, and by the unusually high pressure in the Northwest of the United States at the time of cold waves, it is possible that a thin stratum of the lower very cold air travels entirely across to the Eastern States; and this, when the temperature of the earth's surface is very low, and especially when it is covered with snow, which prevents the heat of the earth below from coming through readily to warm up the air above, continues very cold all the way. Under these conditions the flow of air from the north or northwest may extend to a long distance before its temperature is much changed. The coldest winter weather is experienced in the middle and southern latitudes of the United States, when the whole country is covered with snow, and there is an area of very high barometer and low temperature in the northern part, or in British America. But whether the air in the rear of a cyclone comes from the extreme Northwest, or directly down from higher latitudes, we are not safe in predicting a cold wave in the middle and southern latitudes, from the occurrence of an area of very high barometer and low temperature in the North or Northwest; if the intervening surface of the earth is warm and uncovered with snow, since the stratum of cold air becomes warmed up before it travels very far.

The cold waves are said to not occur in summer, but even in this season there are similar changes, but of course not so marked, which give rise to agreeable changes from very warm to cooler weather, and which may be called "Cool Waves."

As most of the areas of low barometer, or, in other words, cyclones, originate in the region east of the Rocky Mountains, so the cold waves, which are their rear parts, we have seen, originate there also, and both progress across to the Atlantic Ocean together.

Again Lieut. Woodruff says:

"It is observed that some of the severest 'northers' in Texas occur when an area of low barometer appears near the coast, the winds at Galveston and Indianola are easterly, and the temperature high; the 'low' moves northeasterly, the winds back to northerly and northwesterly, and the temperature falls more or less rapidly, according to the rapidity

with which the low moves off. If the storm have considerable energy, the 'norther' is severe and sudden."

Here again the cold wave or norther is the west side of a cyclone, and the intensity of the former seems to depend upon that of the latter. The northers of Texas are noted for the greatness and suddenness of the changes from high to low temperatures. The cyclone here which gives rise to the northers being near the coast of the Gulf of Mexico, and the contrast in winter between the temperature of the land and the Gulf being very great, the warm air of the Gulf is carried northward over the land on the east side of the cyclone, while the colder air of higher latitudes is brought down on the west side, and pressing under the cyclone, comes in contact with, and rather meets, the southeasterly much warmer and moister air on the other side; and when this line of meeting, or trough of the cyclone, passes over a place, there is a great and sudden change of temperature.

215. An interesting and instructive paper on the northers of Texas was read at the meeting of the American Association for the Advancement of Science about 20 years ago by Solomon Sias,[58] who seems to have resided in Texas for some time and to have made many observations of them. The following extracts are taken from this paper and given here with the explanations and comments following:

"The wind from whatever quarter blowing, usually S.S.E. or S.W., either entirely dies away or very materially slackens, and is changed to a cold, piercing north wind. This is a norther. . . . Frequently, we may say usually, the change is so sudden and marked that a person standing in the open air feels it slap him with a chilling roughness, and almost immediately the moisture is dried upon him which the preceding warmth had produced. While riding over the prairies, uncomfortably warm in the lightest clothing, I have repeatedly been struck by them, and before I could wrap my blanket around me, been as uncomfortably cold."

"Sometimes, instead of changing, the preceding wind dies entirely away, and a dead, oppressive, suffocating calm ensues, to be broken in a few hours by the wild bursts of the descending norther."

"They frequently commence faint as a summer's zephyr, again bend the trees like reeds, and I have known the brick walls of our Institute to

quiver at the first striking of the blast. . . . They die down in a few hours to a force of about two, hold at this rate perhaps a day or so, then fade entirely away."

"The wind in a norther is not always strictly from the north; it frequently veers for a few hours, to N.E. or N.W., or back and forth between these points, and I have known it to give way completely for an hour or two to a southerly wind. This veering and changing, however, seldom occurs in the early stages, or in a norther of a high degree."

"The total number of northers in a year varies from thirty-five to forty-eight; the average is about forty-two or forty-three. We are told they commence the last of September and end in May; but meteorological observations show they commence earlier and end later; in fact, there is no month in which they may not occur."

"The thermometer frequently falls rapidly at their commencement, —it is said, sometimes seventy degrees in fifteen minutes; but I have never witnessed such rapid or extreme falls. The greatest I have noted is twenty degrees the first hour and fifteen the next, making thirty-five degrees in two hours; and this is a very exceptional case."

"Usually the barometer commences falling from two to six days before a norther sets in, and drops down slowly but pretty regularly until the first stroke of the norther, when it rises rapidly. Frequently the fall is more rapid just before the change; and I have often been led to the belief that a norther was close at hand by this phenomenon, in the absence of other usual indications, and I do not remember ever being disappointed. And the almost invariable fact that it rises the moment one begins has sometimes been my first and surest evidence that one is blowing."

"The northers may be divided into two classes,—the wet, or those accompanied by rain, sleet, or snow; and the dry, in which the sky is clear, or but partially covered with clouds. If the preceding wind has been east or south, we usually look for a wet norther; if it has been directly south, the sky laden with clouds, and the norther does not scatter them immediately, it may be wet; if the wind has been west of south, it is usually a dry norther."

"The northers are mere surface winds. When the wind is ranging from three to five, the clouds, and even down, are frequently seen floating in the opposite direction. The gusts of wind sometimes seem to actually roll along the ground."

As heralds of the approach and attendants during the progress of a norther, Mr. Sias states:

"A warm, moist wind blows from some southerly quarter a few days; the thermometer rises; the barometer sinks slowly, then rapidly; the

wind materially slackens, veers to the west or gives way to a dead, oppressive calm; and lastly, a peculiar dark cloud-like appearance forms in the northwestern horizon, slowly rises, and when in a few hours it reaches an angle of thirty or forty degees, the norther bursts upon us."

"The attendants are usually the immediate rise of the barometer and falling of the thermometer; sometimes a dash of rain, occasionally the ozonic smell and curdling of the air; almost invariably the rapid disappearance of the northern cloud-like formation; and frequently so great a reduction of temperature that a frost or freezing occurs out of season."

216. From a mere reading of these extracts and those from Lieut. Woodruff's paper, and comparing them with § 193, it is seen that in cold waves and northers we simply have the usual trough-phenomena of cyclones in their passage over a place, where these phenomena are well marked. These, we have seen, occur mostly in winter, and in latitudes where cyclones have an easterly progressive motion, and rarely in summer; and precisely the same is true of cold waves and northers. In all there are, first, mostly southerly and southeasterly, warm and damp winds, accompanied by a gradually falling barometer, which toward the last becomes very rapid. Then comes the dividing line between the warm southerly and southeasterly, and the cold northwesterly winds, which, on account of the great difference of temperature, do not readily mix, and so there is a sudden passage from the one to the other. There is, on this line, a very steep pressure gradient for a short distance, which, however, soon passes over, and during this time the squalls are often terrific; after which there are, for several days, the usual westerly winds of the rear of the cyclone. Sometimes, however, it seems that there is not this very sudden change, but the central dead calm is observed for a short time. This is, no doubt, when the centre of the cyclone passes over the place, and when, for some reason, the trough-phenomena are not so well developed as usual.

The northers, like cyclones, may or may not be accompanied by rain, that is, at any given place, though there is, perhaps, always more or less rain somewhere. When the preceding winds are west of south, as they are mostly when the track of the cyclone is on the north side and the place is on the south

side of the cyclone and out of the rain area, there is no rain, and the norther is called a dry norther. On the other hand, if the preceding winds are southeasterly, as they are when the central part or the northern side of the cyclone passes over the place, and this falls within the rain area, the norther is called a wet norther.

The norther is said to be a mere surface wind, and the clouds above are frequently seen floating in the opposite direction. A stratum of the cold air of no great depth is naturally forced under the warmer and lighter air on the southeast side of the cyclone, where the usual cyclonic motions, giving rise to southerly and southeasterly winds, prevail. Besides, at the dividing line, where the pressure gradient is very steep for a little distance, and the air rushes out with great velocity, there is a return current above at no great height.

At the line of meeting of the winds of nearly opposite directions and great contrasts of temperature, the warm moist winds are thrown up, and somewhat over the stratum of cold air which they meet, and the cooling of this moist air, both by ascent and by coming into contact with the cold air, causes the usual "dark, cloud-like appearance which forms in the northwestern horizon," and which is usually seen at some distance, and for some time before the norther arrives, but which generally soon passes by. In high latitudes, where the colder air is below the freezing-point, the mingling of the warm and moist air with it gives rise not only to a very strong wind, but also to a dense snow-storm, called a "blizzard," * in which persons who are caught are sometimes unable to find their way to a place of shelter and protection, and often perish.

Since cold waves, by the preceding explanation, depend upon cyclonic gyration and a large temperature gradient in a north and south direction, they should not only abound mostly in winter when the gradient at any place is largest, but also in countries where, at the same season, this gradient is greatest.

* This term is said to have had its origin amongst the Germans of Dakota, who called a wind of this sort a *blitzartig* (lightning-like) wind, and finally a *blitzart* simply, which was changed by the English settlers to *blizzard*.

In the Mississippi valley, between the warm Gulf of Mexico and the cold regions of the northern and northwestern part of North America, the temperature gradient is very steep. And as the changes depend upon the progressive motions of the cyclones, these changes should be most sudden where the progressive motions are most rapid, as in the United States of America. Hence here the cold waves are most numerous, and the changes of temperature unusually great and sudden. This seems to be the view taken of the matter by Dr. Hinrichs, who says (*Am. Met. Journal*, Feb., 1888):

"In winter the great thermic gradient of the Mississippi valley is the real cause of the sudden changes in our weather. In this thermic contrast between the north and the south we have the origin of our blizzards and cold waves. This great difference in temperature makes it possible for us to have a thunder-storm in winter, followed in a few hours by a blizzard. The irregular changes of temperature from day to day are accordingly much greater in the Mississippi valley than in Russia. Our so-called cyclonic storms travel about twice as fast as those in Europe and Russia."

PAMPEROS.

217. The pamperos of South America, cold southwest winds from the Pampas, or plains, are the same as the cold waves and northers of North America. This will be at once seen from a few extracts from Dr. David Christison's paper on the pamperos of Central Uruguay:"

"The barometer fell pretty steadily from two to four and one half days before the storms, the fall varying from 0.18 to 0.56 of an inch. In two it began to rise some hours before the storm burst, and it may possibly have done so in all. In all it continued to rise for some days after the pampero.

"The *thermometer* in all cases showed a marked tendency to rise for some days before the pampero. Unless interrupted by the occurrence of rain, thunder, or a temporary shifting of the wind to a cold southern point, this rise seemed to go on pretty steadily, a few degrees being added every day to the heat, in some instances for more than a week. After the pampero the fall was rapid, but the temperature soon began to rise again, particularly with a change of wind to the east. The coolness produced by the pampero is one of their most striking characteristics,

and although it occurs at all seasons, it is particularly welcome in hot weather. It happened in every one of the twelve cases under consideration. Comparing the daily maxima before and after them, the average fall of temperature was 13° F., the least being 4°, and the greatest 24°. The greatest absolute fall was 44° in 14 hours; in another instance it was 33° in 6 hours.

"*Thunder* accompanied seven, and immediately preceded two others of the twelve. It also occurred within a few days before or after several of them; but three were not accompanied with thunder at all.

"*Rain* accompanied eight, immediately preceded two, and closely followed another. In some instances it was very heavy, in others quite trifling. There was also a good deal of rain within a few days before them, and a small amount within the same period after them. Only one was quite unconnected with rain.

"*The wind* before this class of pamperos almost invariably blew moderately or gently for several days from easterly points, perhaps shifting to the N.; the change to S.W. sometimes takes place from the latter quarter. Any deviation to other points seemed to be quite exceptional. The change to S.W. was sometimes preceded by a calm; but in most cases that I had the opportunity of watching closely, the east or north wind continued to blow, although with diminished force, until the moment when the pampero supplanted it. For some time before this, however, the clouds overhead were either motionless, or moving very slowly from the N.W., more rarely from the N. or W.

"Perhaps the most striking characteristic of this class of pamperos was the invariably sudden outburst of the wind at its full strength almost from the first. Continuing to blow thus steadily from ten to thirty minutes only—in one case for an hour—it then either ceased entirely, or more frequently continued with diminished force for a certain number of hours. The force of these short outbursts varied from that of a strong gale to a moderate puff of wind, but was always sufficient to mark them off distinctly from the preceding or following wind.

"Sometimes the wind continued in the S.W. for only twelve hours, but never more than three days, before returning to some easterly point. In general it spent the greater part of the time in the S., changing to that quarter soon after the pampero had blown itself out. But in some cases it changed several times between S.W. and S. before reverting to the east."

From these extracts it is evident that the sequence of phenomena preceding and attending pamperos are the same as in the case of cold waves and northers, except that in the directions of the winds, on account of their being in the opposite

hemisphere, we must put N. for S. and S. for N. There is first, for several days, the gradual falling of the barometer and increase of temperature with damp winds from E. to N.; then, sometimes a short calm, but mostly a sudden change of the wind to S.W., which is the pampero proper. This at first is generally very strong, and then more moderate, and is accompanied by a rapid lowering of the temperature and increase of barometric pressure. But the principal characteristic, as in cold waves and northers, is "the invariably sudden outburst of the wind at its full strength, almost at the first." They were usually preceded by several days of rain with thunder, as the trough of a cyclone is, and also followed by some rain.

THE MISTRAL AND THE BORA.

218. Along the whole northern coast of the Mediterranean in winter, unusually cold northerly winds frequently prevail for several days, called in the Rhone valley and the Gulf of Lyons the *Mistral*, in the Adriatic a *Bora*, and in the Grecian Archipelago a *Tramontana negra*, or Black Norther. These winds are always connected with, and form part of, a cyclone, and are the same as the northers of Texas. They prevail whenever there is a cyclone so situated in the Mediterranean that the currents of the northwestern side of the cyclone bring the comparatively very cold air from higher latitudes down to the Mediterranean coast; for along this coast in winter there is a contrast between the land and sea temperature similar to that between Texas and the Gulf of Mexico. But the mistral and the bora are strengthened by the prevailing belt of high pressure north of them in winter, extending from the Atlantic over the Spanish peninsula, France and Austria, on toward the interior of Asia, Wœikoff's "Great Axis of the Continent,"[88] which gives rise to a prevalence of northerly winds at all times along the northern coast of the Mediterranean during the winter. These winds, as those of the cold waves and northers of North America, are also strengthened by any area of high pressure existing from any cause at the time to the north or

northwest of them, and which may be nearly or quite independent of the cyclone.

These winds are especially cold after there has been high pressure and very clear and cold weather over the mountainous regions to the north, when the air, very cold from radiation from the mountain sides, runs down and settles in the valleys. When all the conditions favorable to the norther are at hand, this very cold air is drawn down to the Mediterranean coast, and even on to much lower latitudes. "In the eastern part of the Mediterranean, the northerly winds reach to the African coast, continue over Egypt and the Nubian desert, and have even been felt on the Nile as far south as 8° or 9° N."[11]

Since these winds are always connected with a cyclone, they last generally only a few days, until after the cyclone, in its progressive motion, has passed away. There is here, however, usually no sudden transition from very warm and moist to very cold and dry air, as in the case of cold waves and northers, since the direction of progressive motion is toward the N.N.E., very nearly in the direction of the trough of the cyclone, or line separating the warm winds on the southeasterly side from the cold winds of the northwesterly side.

THE FOEHN AND THE CHINOOKS.

219. It sometimes happens that near the base of a high mountain range for several days there are unusually warm and dry winds coming from the direction of the mountain. These are noted winds on both sides of the Alps, especially on the north side, and are called there and elsewhere the *Foehn;* but in the northwest part of North America, east of the Rocky Mountains, *Chinooks.* Although the true explanation of these winds was given by Espy, yet it is only somewhat recently, since the full explanation of them by Dr. Hann, that they have been generally understood. They are connected with, and are a part of, cyclones, just as the northers of Texas and the Mediterranean.

The effects of several permanent foehns, depending upon the general motions of the atmosphere over mountain ranges, have already been referred to and explained (§ 128). In the same manner temporary foehns are produced wherever there is a cyclone so situated as to draw a current of air over a high mountain range, and the effect is particularly marked where this air is drawn from warmer latitudes and is very moist. For then the air is warm at the beginning of the ascent, and being very damp, it is seen from Table III that after having ascended to only a very moderate altitude, its rate of cooling with increase of altitude in its ascent is comparatively slow, so that on arriving at the top it still has a high temperature for the altitude. In its descent, now, it becomes warmed up 1° C. for each 100 meters, and so on arriving at the base of the mountain on the other side, as it is drawn in toward and around the centre of the cyclone, it has a much warmer temperature than that of the surrounding air, and than that which usually prevails at the season. And having lost most of its vapor in ascending to the top of the mountain, it of course becomes a very dry wind after descending to the base of the mountain on the other side, down perhaps to about the same level as that from which it started.

Dr. Hann has shown that the foehn of Switzerland occurs when there is a large cyclone of considerable intensity with its central depression in the direction of the British Isles, or beyond in the Atlantic Ocean, and high barometer S.E. of the Alps, and the foehn on the south side of the Alps, when the reverse is the case; that is, when the low pressure is in the Mediterranean S.E. of the Alps and high pressure in the direction of the British Isles. He has given a graphic representation of each, which may be regarded as typical cases—the one of the great foehn of January 31, 1885, south foehn in Switzerland, and the other of the foehn of October 5, 1884, north foehn, on the south side of the Alps.[60] In the case of the south foehn, the wind is drawn across the Alps in the southeast part of the great cyclone, and in the other from the north side by the slight cyclonic action in this case of a cyclone

in the Mediterranean, but mostly forced across by the very high pressure in the direction of the British Isles. The foehn on the north side is generally the most marked, because of the permanent cyclone in the winter season in the north Atlantic; for the depression of an ordinary progressive cyclone, added to that of the southeastern side of the permanent one, frequently causes a great depression on the eastern coast of Europe or the contiguous part of the Atlantic.

220. The chinook winds of Virginia City, of our Northwest Territory, have been investigated by Professor Harrington" from the bulletins of Weather Reports of the Signal Service for several years. He shows that they are of the same character as the foehns of Switzerland and other places, and defines them to be "*warm, dry, westerly or northerly winds occurring on the eastern slopes of the mountains of the Northwest, beginning at any hour of the day, and continuing from a few hours to several days.*" Again he says: "*The chinooks occur when a cyclone is passing to the north of the place of observation.*" This condition is necessary in order that the westerly wind of the southwesterly part of the cyclone may draw the air across the main divide of the Rocky Mountains to the place of observation on the east side.

The chinooks are felt, more or less, along the whole eastern side of the Rocky Mountain range, from the southern part of Colorado at least as far north as Peace River, in British America, and to a considerable distance east of the range, the area in which they are most marked including most of Wyoming, much of Montana, and some of Dakota, and they are perceptible as far east as Bismarck, about 500 miles east of the main divide. "A marked effect of the chinooks is the amelioration of the winter temperature caused by them. On the arrival of the chinooks the winter appears to yield. The air becomes mild, and to the residents of this naturally dry region appears balmy and spring-like." "

To the effect of the frequent occurrence of chinooks in this region, but perhaps mostly to the more permanent foehn effect of the westerly winds due to the general motions of the atmos-

phere, as explained in § 128, is due the general mildness and dryness of the winter in comparison with the greater severity and dampness experienced farther toward the east on the same latitudes. So dry are these warm winds that they immediately evaporate the melted snow, and it seems to pass directly and very rapidly into vapor without wetting and softening that which is left. They are consequently sometimes called "snow-eaters."

With regard to the chinooks of British America, George M. Dawson[92] of Ottawa, Canada, says:

"As experienced, the chinook is a strong westerly wind, becoming almost a gale, which blows from the direction of the mountains out across the adjacent plains. It is extremely dry, and, as compared with the winter temperature, warm. Such winds occur at irregular intervals during the winter, and are also not infrequent during the summer, but being cool as compared with the average summer temperature, are in consequence then not commonly recognized by the same name. When the ground is covered with snow, the effect of the winds in its removal is marvellous, as owing to the extremely desiccated condition of the air, the snow may be said to vanish rather than to melt the moisture being licked up as soon as it is produced."

Similar foehn-like winds are experienced in all parts of the world, under different names, wherever a current of air is drawn or forced over a high mountain range. They are more marked in the winter season of high latitudes, because in this case the vertical gradient of the undisturbed atmosphere is small, so that the temperature of the air, descending from a high altitude and heated at the rate of 1° C. for each 100 meters, is much higher at the lower level than that of the surrounding air, or the air generally over the region. In the summer season during the day when the earth's surface is very warm, and the decrease of temperature with increase of altitude great, the foehn effect is small, and the conditions may even be such as to give them a lower temperature than that of the surface air generally.

THE SIROCCO.

221. There is in Italy and the eastern part of the Mediterranean still another warm wind which is connected with cyclones, and is the counterpart of the Mistral and the Bora—*the Sirocco.* It is a very warm, and generally damp, south and southeast wind on the eastern side of a cyclone, with its centre westward of the place of observation. The very warm and dry air from the Sahara and the northern coast of Africa is whirled around toward the northeast and north, and in passing over the sea it becomes a very damp as well as warm wind, and as such mostly it is experienced in Sicily, Italy, and the eastern parts generally of the Mediterranean and the adjacent coasts. Although the temperature never rises above 95° F., yet on account of the great dampness these winds are very oppressive, and seem to be of a higher temperature. This is due to local circumstances, to the existence of the highly heated Sahara from which warm air is drawn, which in its passage over the sea becomes nearly saturated with vapor; but the same in some measure is observed in nearly all parts of the world on the eastern sides of cyclones, where much warmer and damper winds than usual prevail. Like all phenomena of this kind connected with and dependent upon cyclones, they continue only a few days, until the cyclone has passed away. The dust raised from the Sahara and carried northward by the sirocco often falls over the countries north of the Mediterranean as "blood rain" or as "red snow," the moisture and the sand falling together in the rain or snow.

As in the case of the Mistral and Bora, there is generally no sudden transition from very warm and moist air to very cold and dry air, as in the case of the northers of Texas, since the direction of the storm's path here is nearly that of the trough of the cyclone, which separates the winds of the two characters. There is, however, always the same marked contrast between the two sides of the storm, and the storm's path may sometimes have such a direction as to give a considerable change of air in the passage of the storm.

Dr. Hann[60] has given a graphic representation of the great Sirocco storm of the 25th of February, 1879, in the Adriatic Sea, and of its path, which lay over the northwestern part of Italy, and extended in a direction N.N.E.

CYCLONES WITH A COLD CENTRE.

222. If for any reason the central part of any given portion of the atmosphere of a somewhat circular form is maintained in any way at a lower temperature than the surrounding parts, and the temperature gradient on all sides is somewhat symmetrical, we have approximately the conditions which give rise to a cyclone. In this case it is readily seen that there must be a vertical circulation as in the ordinary cyclone, but that it is reversed, out from the centre below, and in toward the centre above, with a gradual settling down of the air in the interior to supply the outward current beneath. This vertical circulation, as in the case of the ordinary cyclone, gives rise to a cyclonic motion in the interior and an anti-cyclonic in the exterior part of the air under consideration, but in this case the gyratory velocity is greatest above and is less at lower altitudes, diminishing down to the earth's surface, where it is least. In the anti-cyclonic part the reverse takes place, the gyratory velocity being least above and greatest down near the earth's surface. The distance from the centre at which the gyratory velocity vanishes and changes sign, is greatest above and gradually becomes less, with decrease of altitude down to the surface, where it is nearest the centre. Consequently, contrary to what takes place in an ordinary cyclone, the upper part of the air, taken over the whole area, is mostly cyclonic, especially if the friction at the earth's surface is but little, and there is considerable gyratory velocity below in the cyclonic part; for, whatever this is, there is always a certain difference between the gyratory velocities above and below, greater or less according to the amount of temperature gradient between the central and exterior parts. If the atmosphere were in the unstable state for dry air, and for any reason it should receive a down-

ward motion over a given area, either from a lower temperature or for any other reason, then this motion would continue, and the central area would continue to be colder as long as this unstable state continued, and during this time the vertical circulation would give rise to and maintain a cyclone with a cold centre, which would be a progressive one. But for reasons given in § 162, we have reason to think the atmosphere is never reduced to this state, except through a stratum of inconsiderable depth next the earth's surface. Hence we have no progressive cyclones with a cold centre.

The cases in nature which give even rough approximations to these conditions are few, and these are such as to give rise to stationary cyclones only. An island, such as Iceland, in winter, when the air over the whole island, and especially the interior part, is much colder than that of the surrounding ocean, furnishes approximate conditions which must give rise to considerable cyclonic action of this sort, though it is all included in the very much larger ordinary and stationary cyclone of the North Atlantic in the winter season.

223. The conditions of a cyclone with a cold centre which are the most nearly perfect are those furnished by each hemisphere of the globe, as divided by the equator, in which the pole is the cold centre and the temperature gradient from the pole toward the equator is somewhat symmetrical in all directions from the centre. It is true, the deflecting force of the earth's rotation is greatest at the centre and decreases with increase of distance from the centre, and vanishes at the equator, and the curvature of the surface makes the conditions vary considerably from those of a small portion of the earth's surface, but still the general results are the same, and all the explanations given with regard to the vertical circulation and the east and west gyratory motions in the case of the general motions of the earth are applicable to the effects of cyclonic conditions of this sort over a small area of the earth's surface. The easterly motions in the higher latitudes and the westerly ones in the lower latitudes, in the one case, correspond to the cyclonic in the interior and the anti-cyclonic in the exterior

part, and the belt of high pressure near the tropics to that of high pressure in the case of any cyclone with a cold centre.

Looking at the general circulation of the two hemispheres of the globe as two cyclones of this sort, we see in another light the cause of the annual shifting of the equatorial and tropical calm-belts north and south, and why they are a little nearer the north than the south pole. The tendency of the two cyclones is to expand over a greater area, and the more so the greater the internal energy upon which their actions depend and the less the amount of friction at the earth's surface. As the outer limit of each is common and the one tends to encroach on the other, during the winter of the northern hemisphere, when the temperature gradient between the equator and the pole, and consequently the energy of the cyclone, is greatest, while at the same time that of the southern cyclone is least, the northern cyclone encroaches a little on the territory of the southern one, and consequently becomes larger, while the other becomes smaller. Hence at this season all these calm-belts have a position a little farther south than their mean position. During the summer of the northern hemisphere, on the contrary, the energy of the northern cyclone is diminished and that of the southern increased, and consequently the reverse takes place, and the calm-belts now have a position a little north of their mean position.

The less, also, the frictional resistances to the gyratory motions of a cyclone, the greater is its tendency to spread and encroach upon the territory of an opposing one, and hence the southern cyclone, being mostly on an ocean surface, is one of much greater violence with the same amount of energy, and the depression at the south pole is much greater than that at the north pole, and the tendency is to occupy a little more territory than its opposing less violent cyclone, and so the mean position of all the calm-belts is a little farther from the south than the north pole.

224. The centre of a cyclone with a cold centre may, or may not, have a minimum pressure, according to circumstances. A certain amount of temperature gradient, and of pressure gra-

dient which is independent of the gyratory motion, as explained in § 72 in the case of the general circulation of the atmosphere, is necessary to overcome the friction in the lower strata and to keep up the vertical circulation, upon which the cyclonic depends; and the pressure gradient, which depends upon the temperature gradient and is independent of the gyrations, may be such that the increase of pressure in the central part due to this cause may be greater than the decrease of pressure arising from the cyclonic gyrations, especially where surface friction is great. For instance, in the southern hemispherical cyclone, where this friction is comparatively small, the cyclonic gyrations are comparatively large, and the effect is to diminish the pressure there more than it is increased by the colder and heavier air there. And this is also in some measure the case in the northern hemispherical cyclone; but with still a little more of land surface, and with still greater mountain ranges, the resistance to the gyratory motion might be such that the decrease of pressure at the pole from this cause would not be equal to its increase from the lower temperature and greater density of the air.

Each of the great continents of Europe, Asia and of North America furnish, in winter, only very roughly the conditions of a stationary cyclone with a cold centre, as they do in summer those of an ordinary cyclone, both because the areas are too large and irregular, and also because the deflecting forces are very different on the polar and equatorial sides, and consequently not symmetrical in all directions. The surfaces also are very uneven, and consequently the resistances to gyratory motion great. There is, however, some motion of this sort, no doubt, but it is not sufficient to reduce the pressure in the central part as much as it is increased by the greater coldness and density of the air in winter, for the effect of this is very great, and so there is a maximum pressure in the interior of these continents during the winter. The whole effect of the unequal distribution of temperature, therefore, has been treated in Chapter V, and regarded as a winter monsoon, some allow-

ance, however, being made for the deflecting force of the earth's rotation upon the directions of the monsoon winds.

225. Regarding the general motions of each hemisphere as a cyclone, then all ordinary cyclones become secondaries, and their effects upon the isobars and the general motions of the atmosphere become similar to those of the secondaries of ordinary cyclones, as stated in § 208 and represented in Fig. 13. If the depression of the cyclone is considerable, so that its pressure gradient is greater than that of the gradient between the equator and the pole where the cyclone exists, then the cyclone gives rise to another minimum, just as in the case of the great stationary winter cyclone of the North Atlantic, as represented by the right-hand secondary minimum in Fig. 13; and the winds gyrate around this centre, but with a velocity much greater on the equatorial than on the polar side, in the middle and higher latitudes, and with a correspondingly steeper gradient. In fact, we have here the resultant of two motions in nearly the same direction, as explained in § 202, and it becomes the dangerous side of the storm. In the northern hemisphere the general gradients between the pole and the equator at all times are so small that any cyclone of only a very moderate barometric depression gives a minimum, but in the southern hemisphere, on the middle latitudes, as in the region of Cape Horn, where the general hemispherical gradients are very steep, unless the cyclone has a considerable barometric depression of its own, it does not give a minimum, but simply causes a derangement in the general isobars extending east and west around the globe, making them closer on the equatorial side, and the isobars on the other side do not inclose a minimum, but are open, as represented in the left-hand secondary in Fig. 13. However, with a little greater cyclonic depression, there is a minimum even here, as represented by the right-hand secondary in Fig. 13.

It is readily seen, therefore, that the minimum pressure in an ordinary cyclone, as observed and laid down on synoptic charts, being the resultant of two separate effects, is not the place of the true centre, especially in the southern hemisphere,

where the general gradient is steep. In fact, without a very marked cyclonic depression, the cyclone centre would be indeterminate, since there is no minimum of pressure, or isobars inclosing an area of lower pressure. In tracing the paths of cyclone centres, therefore, unless there is a marked cyclonic depression, there is always considerable error from not taking these considerations into account, and the error in the middle latitudes of the southern hemisphere would generally be very great.

226. The great permanent cyclone of the North Atlantic in winter may be regarded as a secondary cyclone with reference to the great hemispherical cyclone in which it is situated, and as the ring of high pressure of the secondary on the equatorial side falls somewhat upon that of the primary, that is, upon the tropical belt of high pressure around the globe, the effect is similar to that represented in Fig. 13, in which the resultant, where the two rings fall together, is a pressure considerably higher than elsewhere in the ring of high pressure of the primary. This is the explanation in part of the area of unusually high pressure in the Atlantic between Spain and the United States. The other part arises from the gyration of air around this region resulting from the deflections of the coasts and mountain ranges, as explained in § 123. The deflecting force arising from the earth's rotation being on all sides to the right, and so in toward the centre of this region, causes a little accumulation of atmosphere and increase of pressure.

AREAS OF HIGH BAROMETRIC PRESSURE.

227. If only a single regular cyclone, without any other abnormal disturbances, existed in any part of the globe, in which the atmosphere in its normal condition had a uniform barometric pressure at all places, we have seen, § 174, that there must necessarily be a ring of barometric pressure around the central part of the area of low barometer, a little above the normal mean pressure. But it has been shown that the pressure of the atmosphere, undisturbed by transient and progress-

ive cyclones, is not of uniform pressure at all places on the earth's surface, but is made to vary in different latitudes by the general motions of the atmosphere, and both in latitude and longitude by the stationary cyclones, and unequal distributions of temperature between land and water, both in winter and summer, giving rise to monsoons, if not regular and stationary cyclones. If, therefore, the inequalities of pressure of regular cyclones are superimposed upon all these other irregularities, a chart of the resultant pressures does not give regular circular isobars, and indicate a regular ring of high pressure around an area of low pressure, but the former are very much distorted, and the latter is broken up into areas of higher and lower pressure. The duration of the area of high pressure depends upon that of the cyclone, and generally has an easterly progressive motion, somewhat as that of the cyclone, but not necessarily the same, since it depends upon many other irregularities upon which it may be superimposed.

But two or more progressive cyclones may interfere with and overlap one another, and then the irregularities thus produced, together with other more permanent irregularities, may make the barometric pressure very irregular and cause great distortions in the isobars of the charts. Across the United States there is generally a pretty regular series of cyclones passing from west to east, at intervals of a few days, of which the part of the ring of high barometer on the west side of the one which precedes, falls somewhat upon that of the east side of the one which follows, causing a sort of ridge of high pressure between them; and if this is still interfered with by other irregularities, as it usually is, it may be an area of high barometer of almost any form.

In forty-four cases of ridges or areas of high barometer with an area of low barometer between, passing over the United States, Loomis[31] found the average distance from the centre of low barometer to that of the areas of high barometer preceding and following to be about 1000 miles, and the average height of the barometer about 30.35 inches, the normal height being about 30 inches. This indicates that the average

height of the rings of high barometer was about 0.2 inch above the normal height, supposing that the highest parts of each did not in general fall exactly together.

228. The areas of greatest high pressure do not generally depend directly upon cyclones, but only indirectly, and directly upon low temperature. On the clearing-up side of a cyclone which has passed over any region, the air is clear and the terrestrial radiation into space great, so that the air becomes unusually cold and dense, and consequently the barometric pressure greater than at surrounding places where the air is warmer. Thus on the morning of the thirteenth of February, 1888, according to the Signal Service Charts, in the northwestern part of the United States, the barometric pressure was below 30 inches, the lowest 29.6 inches, with partially cloudy weather, and temperature little below freezing. By the next morning the low pressure and cloudy area had passed off toward the east, fair and clear weather prevailed, the temperature over that region had fallen about 30° F., and the maximum barometric pressure in Dakota was 30.9 inches. This increase of pressure was evidently due mostly to the lowering of the temperature by radiation into space through the now clear atmosphere. By the morning of the 15th the region of highest pressure, now 31.0 inches, had moved eastward to Lake Superior, the weather being still very clear and cold. After this it still moved farther eastward but rapidly subsided.

The principal cause of the large areas of very high barometer which frequently occur in the higher latitudes in winter is undoubtedly found in the clearness of the atmosphere over these areas and the intense coldness produced by the radiation of heat at a time when little is received from solar radiation. The density and pressure of the air are much increased from this cause, and the areas are too large and irregular for this temperature disturbance to give rise to a cyclone with a cold centre, by the gyrations of which this pressure would be diminished in the central part, as it is in the great hemispherical cyclones, especially that of the southern hemisphere. If there were no gyrations around the poles of the earth, the polar re-

gions, instead of being areas of low pressure, as they are, would be areas of very high pressure.

As there is a gradual settling down of the air in the regions of unusually high pressure, they are clear areas, and since the dry atmosphere has comparatively little radiating power, the cooling takes place mostly at the earth's surface, the contiguous stratum being cooled mostly from contact with the earth's surface, while the air at some distance above is becoming warmed from gradually descending and coming under greater pressure, an effect which is not felt near the earth's surface, since the descending air does not reach it but is deflected laterally and the air near the surface remains under the same pressure. Hence the cooling is so much greater below than at some height above the earth's surface, that the vertical temperature gradient after a continuance for some time of high barometric pressure becomes inverted, and the temperature is higher above than below. A remarkable instance of this kind occurred in France during two periods of January, 1880. The following table shows the minimum temperature observed during these periods at the several places named:[44]

DATES.	Parc St. Maur, Paris (altitude 46 m.).	Poitiers (altitude 117 m.).	Clermont (altitude 407 m.).	Puy-de-Dôme (altitude 1,467 m.).	Pic du Midi (altitude 2,366 m.).
4	−1°.4	−0°.4	−4°.0	−3°.0	0°.2
5	−1.5	−0.6	−6.0	1.0	0.0
6	−1.1	−0.3	−4.0	−2.0	−3.5
7	−1.5	0.0	−1.4	−1.0	−3.3
8	−3.5	−2.3	−7.0	−1.0	−4.0
9	−2.0	−3.3	−7.0	−2.0	−4.2
10	−3.4	−4.4	−7.0	−2.0	−5.0
11	−0.7	−4.2	−8.0	−1.0	−4.5
12	−4.8	−6.1	−12.0	−2.0	−5.2
13	−7.6	−5.4	−7.0	0.0	−5.4
14	Wanting.	−3.2	−10.0	−6.0	−4.8
25	−7.5	−8.1	−14.0	−7.0	−10.9
26	−9.8	−7.1	−10.0	−5.0	−8.0
27	−9.0	−6.1	−12.0	−2.0	−8.2
28	−11.5	−8.0	−14.0	−1.0	−8.2
29	−11.0	−5.9	−5.0	−1.0	−5.8
30	−5.3	−0.4	−4.0	1.0	−5.2
31	−5.4	−1.7	−4.0	1.0	−5.2

The Puy-de-Dôme is about ten kilometers from Clermont. From this table it is seen that during the first and last parts of the month the temperature at Paris, Poitiers, and Clermont was generally lower than at the top of the Puy-de-Dôme, and frequently less than that on the top of the Pic du Midi.

According to the late Professor Plantamour:

"It happens every year that the temperature on St. Bernard at several hours, or even during several days, of December is higher than at Geneva. But during December of 1879, this anomaly lasted during a longer period of time than usual. The average temperature of St. Bernard was 8°.4 C. higher than at Geneva. Out of the 31 days of the month, only during 14 days was it from 0°.4 to 6°.2 C. lower than at Geneva, while during the 17 days it exceeded this by 2° to 16°.4."

CHAPTER VII.

TORNADOES.

229. In addition to the more general atmospheric disturbances considered in the preceding chapters, comprising the general circulation of the atmosphere, monsoons, and cyclones, there is another and distinct class of disturbances, with their attendant phenomena of waterspouts, hail, thunder, etc., which are very local in character, and occupy at any one time only a very small portion of the earth's surface in comparison with that of a cyclone, but which, over this small area, are generally characterized by far greater violence and destructiveness. These smaller disturbances, though differing from one another in many respects, are all somewhat similar in their general character, and all depend mostly upon the unstable state of the atmosphere. They all have more or less gyratory motion, and may therefore all be included under the general name of Tornado, but in this more extended sense of the term it does not include necessarily the popular idea of great violence.

Tornadoes differ from cyclones mostly in their extent. Both have vertical and gyratory circulations; but while a cyclone may extend over a circular area of one or two thousand miles in diameter, a tornado rarely affects sensibly at any one time such an area of one mile in diameter, and generally very much less. To understand clearly the distinction between a tornado and a cyclone, it is necessary to understand the difference in the conditions which give rise to them. A cyclone, we have seen, requires, in addition to the state of unstable equilibrium for saturated air, such a disturbance in the general equality of temperature over a considerable area that there is a central and somewhat circular area of higher or lower temperature, from which arises a vertical, and consequently a

gyratory, circulation, and the initial extent of the cyclone depends upon that of the initial temperature disturbance. The tornado, we shall see, is independent of any such temperature disturbance which determines the initial extent of the atmospheric disturbance, but simply depends upon conditions which give rise to very local disturbances merely.

THE CONDITIONS OF A TORNADO.

230. The principal condition of a tornado is the unstable state of the atmosphere, from which, with any very slight disturbance, arises a bursting up of the air of the lower strata of the atmosphere through those above, over one or more small spots, somewhat as the vapor of boiling water, which is generated mostly at the bottom of the containing vessel, bursts up through the water above and comes to the surface. But this initial start in the tornado having once taken place, the condition of unstable equilibrium tends to continue the initial motions as long as this state continues, unless the whole system is broken up by great abnormal disturbances and irregularities of the earth's surface. The start being once made, the interior part where the air ascends is kept warmer than the surrounding parts, exactly in the manner explained in the case of cyclones in § 159, and illustrated in the table of that section, and while the unstable state and this warmer central part continue, a vertical circulation is maintained, the air ascending in the interior, flowing out in all directions above and in from all directions below to supply the ascending current, just as in the case of a cyclone; but in this latter the vertical circulation is of great extent horizontally in comparison with its height, while in the tornado the reverse is the case, the vertical extent being generally much greater than the horizontal.

The vertical circulation is the initial stage in the formation of a tornado, and so the tornado cannot originate without the condition of unstable equilibrium which gives rise to a vertical circulation. It is not necessary, however, that this state shall extend from the bottom to the top of the atmosphere, and

such a state, perhaps, never exists; but if it does, rapidly ascending air continues to become warmer than the surrounding air at the same level up to the highest strata, and the difference of temperature there is the greatest. This condition would, of course, give the strongest ascending current and vertical circulation. The ascending current in this case has a maximum velocity at some very high altitude, where the pressure is the same as in the surrounding air at the same altitude, the pressure below this level being less than that of the surrounding air at the same level, on account of its greater temperature and less density. Above this level the pressure is greater than in the surrounding air, on account of the heaping up of the atmosphere, the kinetic energy of the ascending current being gradually changed to the potential energy of increased pressure. But this heaping up of the atmosphere and increase of pressure causes a lateral deflection of the ascending currents in all directions in the upper strata of the atmosphere.

If the unstable state does not extend to a very high altitude, the difference of temperature between the ascending and the surrounding air increases up to this level, and then begins to decrease, until at some higher altitude it vanishes, and then above this it becomes colder than in the surrounding air. The level at which the difference of temperature between the ascending and surrounding air vanishes is an isobaric as well as an isothermic surface. The velocity of the ascending current is increased up to this level; above this the air is colder and the pressure greater than in the surrounding air at the same level, but this does not necessarily extend to the top of the atmosphere. Below that level the temperature is greater and the pressure less in the ascending current than in the surrounding air at the same level, and the greater the difference of pressure, the greater the force by which the kinetic energy is generated, and the whole kinetic energy at any level is the equivalent of the sum of the differences of the forces from some lower stratum of the earth's surface where the difference of pressure and the ascending current begins. Above this level the difference of temperature and of pressure is reversed, the

temperature of the ascending current being the less and the pressure the greater, and the greater the difference of pressure the greater the rate at which the kinetic energy of the ascending current is diminished. The condition of stability of the air here may be such that the temperature is diminished and the pressure increased at such a rate that the ascending velocity and kinetic energy are brought to naught before the top of the atmosphere is reached, the increase of pressure being sufficient to counteract the ascending velocity before the top or a very high altitude is reached. In this case the increase of pressure above deflects the air off laterally in all directions up as high as the ascending current reaches, and the vertical circulation extends up to that altitude only as a direct effect of the conditions; but air of a higher altitude may be brought into the circulation by friction.

In what precedes, a vertical circulation merely is considered, and not the modifying effects of a gyratory circulation, which depress, as will be seen, the isothermic and isobaric surfaces in the interior of the gyrating portion of air; so that in this case the relative temperatures and pressures on the same levels become different.

Again, the air may be in the unstable state in the middle strata, and in the stable state both above and below. It also frequently happens that the unstable state exists above, but near the earth's surface the stable state. This is especially the case where the air is not completely saturated. For in this case the unstable state near the earth's surface requires a decrease of temperature with increase of elevation in the surrounding undisturbed air at a rate greater than 1° C. for each 100 meters, which is rarely found, at least to any considerable altitude. In such a case the vertical circulation does not include the stable strata near the earth's surface, except as they may be acted upon by means of friction.

From what precedes, therefore, the conditions of vertical circulation, and consequently of a tornado, does not require that the unstable state shall extend to all strata of the atmosphere from the earth's surface up to the top.

231. But in order that the vertical circulation may give rise to a gyratory circulation and to much violence in a tornado, another condition is necessary, namely, an initial gyratory motion of the air around the central point where the first ascent of air takes place and toward which the air from all sides is drawn. This may be illustrated by means of the behavior of water which is allowed to run out of a basin through a hole in the bottom. If the water is entirely at rest with reference to the basin, it flows directly in from all sides toward and out at the hole without assuming any gyratory motion; but if the water is not entirely quiet, if there is only the least perceptible disturbance, it is liable to run, in the one direction or the other, into very rapid gyrations near the centre. The direction of the gyration, from right to left or the contrary, depends upon the predominance of the gyrations in the mass generally in the one direction or the other, and does not require that the whole mass shall have an initial gyration in one way. The case of the tornado is similar; but instead of running down, the air of the lower strata runs up through the strata above, where the first ascent takes place, and flows away there in all directions. And if there is no initial gyratory motion of the air, it runs directly toward the centre from all sides, up in the central part and out above, and thus a vertical circulation is inaugurated and maintained, but there is no gyratory motion or much violence. But if there are initial gyrations of the air, however slight, at a distance from the centre, the air as it approaches the centre below runs into a gyration around that centre, and the direction of the gyration is determined upon the same principle as in the case of the water in the basin. It is not necessary that the initial motion of the air shall be that of an absolute gyration around this centre as a fixed point, but only that there shall be a whirl in the atmosphere around some point at no very great distance, such as to cause a relative gyratory motion around this centre, just as every part of the atmosphere, except at the equator, has a gyratory motion relative to any assumed point, in consequence of the rotation of the earth on its axis, though an absolute motion around the pole only.

In a cyclone, we have seen, the gyratory motion arises from the absolute motion of the air and the earth's surface around the centre of the cyclone, and not from initial gyrations which the air has relative to the earth's surface. In a tornado, on account of the smallness of its horizontal dimensions, this effect is comparatively small. For instance, at the distance of 1000 meters from the centre of a tornado on the parallel of 45°, the absolute gyratory velocity relative to this centre, arising from the earth's rotation, is $1000 \times n \sin 45° = 0.05$ of a meter per second, obtaining the value of $n \sin 45°$ from Table V, it being one half of $2n \sin l$ for that latitude. This is probably less than the relative gyratory velocity which the air may have with reference to the centre arising from the numerous whirls into which the air is being continually thrown relatively to the earth's surface; but still it seems it must have something to do with determining the direction of the gyratory motions in tornadoes, since this is generally, if not always, the same as in cyclones, and its effect may have been heretofore underestimated.

Although the horizontal extent of the violent part of a tornado is small, yet the air may be drawn in slowly from a much greater distance than 1000 meters, and so at this distance it may have sufficient gyratory velocity to determine its direction; for, as we shall see, a very small gyratory motion at this distance suffices to give rise to a great gyratory velocity and much violence near the centre.

It is said that in making experiments to ascertain whether the earth's rotation has any sensible effect in determining the manner of gyrations in the flowing of water through a hole in the bottom of a large and shallow basin, it is impossible to have the water so quiet that in running out it does not generally run into a gyration the one way or the other, but that there is no observable preponderance of the gyrations in the direction which would indicate that the earth's rotation has any sensible influence upon the direction of the gyrations. This is to be expected in so small an area; but if the extent

of the basin were as large as the base of a tornado, the result would probably be different.

GYRATORY VELOCITY AND ITS EFFECT ON PRESSURE.

232. In a cyclone the base is so great in comparison with the height, that the whole mass of gyrating air may be regarded as a thin disk, and consequently a large amount of the forces is spent in overcoming the frictional resistances at the earth's surface; and the gyratory velocities with regard to distance from the centre do not at all follow the law which would hold in the case of no friction. But in a tornado the height is so great in comparison with the base that it may be regarded as a gyrating pillar of air, and hence the effect of frictional resistance upon the gyratory velocities, except near the surface, is small, and the law of the gyrations at all altitudes a little above the earth's surface is somewhat the same as in the case of no friction, and follows very nearly the law of a free body in the case of central forces. The general expression of this law has been given in § 41, and is $rw = c$, in which r is the distance of the body from the centre, and w is the absolute gyratory velocity of the body at that distance around that centre, c being some constant quantity. Or if, for simplicity, we neglect the small influence which the earth's rotation may have, it becomes

$$rv = r'v' = c,$$

in which v is the gyratory velocity relative to the earth's surface, and in which v' is the value of v at the distance of r' from the centre. It must be understood here that v is simply the gyratory component of a motion which may also have both a horizontal and vertical component of motion. From this law it is seen that the gyratory component of velocity v is inversely as the distance r from the centre, and hence near the centre it becomes enormously great.

After the vertical circulation in a tornado is fully established and each particle of air has passed in toward the centre and out several times, assuming that no part of the gyrating

column of air is changed, although the initial gyratory velocities at the same distances may be very different at different altitudes and at different distances from the centre, yet by the mutual action of the particles upon one another by contact and by friction, and from the tendency of each one to follow the law of a free body with regard to gyratory velocities at different distances, they all soon have approximately the same gyratory velocities at the same distances from the centre, and nearly follow the preceding law, which makes the gyratory velocities inversely as the distance; especially at a little distance above the earth's surface where the effect of friction is small. The value of c, then, does not depend upon the initial value of v' at any assumed distance r', but upon the average of rv for the whole mass of air brought into circulation.

233. From the preceding law we get

$$v = \frac{r'v'}{r},$$

so that for any assumed value of v' at the distance r' we know the value of v corresponding to the distance r.

The centrifugal force of the gyratory velocity v at the distance r is, putting v for w, § 36, since we assume here that they are sensibly the same,

$$F_c = \frac{v^2}{r}m = \frac{r'^2v'^2}{r^3}m,$$

by putting for v in the first form of expression its value above to obtain the second form.

This expression may be obtained directly from that of F_c in § 40 by neglecting ω in comparison with v, the former, as has just been shown, being very small in tornadoes in comparison with v.

This force, which becomes very great near the centre where r is small, causes a depression of the isobaric surfaces all around the centre of gyration, which becomes very great near the centre. In the figure following let the curved line *beaf* repre-

sent a vertical section of the isobaric surface brought down to the surface of the earth DE at f, and let ed represent a vertical column of air of unit base, which being entirely arbitrary, may be infinitely small, and ad a similar horizontal prism or cylinder of the same base or cross-section. Since the pressure of a fluid is equal in all directions, in order that the air may be in a state of static equilibrium, we must have the horizontal pressure ad in the direction of d, arising from the centrifugal force of the gyration, exactly equal to the vertical pressure of ed at the base d arising from the force of gravity. These spaces being supposed to be very small, the density in all parts is sensibly the same, and using δ to denote this density, we

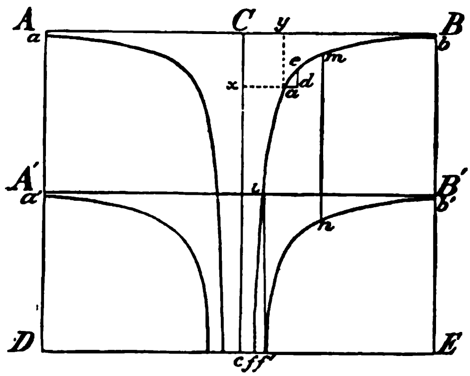


Fig. I.

have $ed \times \delta$ and $ad \times \delta$ for the masses respectively of these horizontal and vertical air columns. Since the vertical force is equal to the acceleration of gravity multiplied into the mass, the expression of this is $g\delta \times ed$. And putting $ad \times \delta$ for m in the preceding expression of the centrifugal force F_c, we get $(r'^2v'^2\delta/r^3) \times ad$ for the expression of the latter in this case. Equating these two expressions, we get

$$g \times ed = \frac{r'^2v'^2}{r^3} \times ad,$$

as a condition which must be satisfied in determining the curved line $beaf$.

Supposing the arc *ae*, Fig. 1, to be very small, and putting, as in § 38, *e* for the ratio between the lines *ed* and *ad*, we get from the preceding equation

$$eg = \frac{r'^2 v'^2}{r^3},$$

in which *e* is the gradient of the curve between the points *a* and *e*, supposed to be very near to each other.

This could have been deduced directly from the expression of *eg* in § 38 by substituting *v* for *w*, since the effect of the earth's rotation here being supposed to be insensible, *v* becomes sensibly equal to *w*, and then substituting for *v* the preceding expression of it. For it is very evident that the gradient of a surface upon which a body, acted upon by the force of gravity and a horizontal centrifugal force, must be the same as that of a liquid surface, or of an isobaric surface, where the liquid is subject to the same forces.

It is seen from the preceding expression, since g, r', and v' are constants, that as r^3 increases, the gradient *e* of the curve diminishes, and at an infinite distance from the centre *C*, Fig. 1, it must vanish,—that is, the isobaric surface must become a level surface. Also, near the centre, where *r* is very small, the gradient becomes very steep, and the more so the nearer the centre.

234. If we let *Cx* and *Cy*, Fig. 1, represent the two rectangular co-ordinates of any point *a* of the curve *beaf*, then $Cy = xa$ is represented by *r* in the notation above, and denoting the line $Cx = ya$, the depression of the curve at the point *a*, by *l*, the equation of the curve or expression of *l* becomes*

$$l = \frac{r'^2 v'^2}{2gr^2}.$$

* By the principles of the differential calculus, if we suppose the points *a* and *e* in the curve, Fig. 1, to be infinitely near to each other, the lines *ed* and *ad* are represented by the differential expressions $-dl$ and dr respectively, and the equation becomes

$$-gdl = r'^2 v'^2 r^{-3} dr.$$

The integration of this gives the expression of *l* above.

From this expression it is seen that neither one of l or r can vanish unless the other becomes infinitely great, and consequently the curve cannot cut either of the lines CB or Cc, in Fig. 1, produced, but simply approximates to them and touches them at an infinite distance only, as a hyperbolic curve does its assymptotes.

For any other isobaric surface which, before being depressed by the centrifugal force, corresponded to the line $A'B'$, either below or above the other, it is evident that we must have a similar curve, since v' is supposed to be the same at all altitudes. If it is lower, it is brought down to the earth's surface represented by the line DE, at a point f' farther from the centre than f, but, if higher, at some point between c and f. But however high the undisturbed isobaric surface may be, the value of r at the earth's surface cannot vanish. Also, the vertical distance mn, Fig. 1, between the two isobaric surfaces, is the same at all distances from the centre, and equal to the vertical distance aa' or bb'.

It is seen from the preceding expression of l that it vanishes only where the value of r is infinite, whatever the value of $r'v'$. But the greater v' is at any distance r', and in the ratio of its square, the greater is the depression l at any given distance r from the centre, until l becomes equal to the height Cc, Fig. 1, of the undisturbed isobaric surface AB; and the value of r corresponding to this value of l in the preceding expression is the value of cf, or the distance from the centre, at which the isobaric surface meets the earth's surface. With this value of r we get from the expression of v, § 233, where the value of v' at the distance of r' is known, the value of v at the distance cf from the centre.

If in the preceding expression of l we assume that at the distance of 1000 meters from the centre of the tornado, as that of D or E, Fig. 1, the gyratory velocity v at all altitudes is 3 meters per second, then we have $r'v' = 1000 \times 3 = 3000$, and

$$lr^2 = \frac{9,000,000}{2 \times 9.806} = 459000.$$

If we now suppose that the height of any undisturbed isobaric surface is 1000 meters, we then get, by putting $l = 1000$, the value of $r = 21.4$ m., for the distance cf, Fig. 1, at which the isobaric surface is brought down to the earth's surface. By assuming any other value of l less than 1000 meters, we get the value of r corresponding to it. For instance, if the depression, as at a in the figure, is $ya = 300$ meters, we then get xa or $r = 39$ meters nearly. In this way a sufficient number of points can be determined to construct the curve.

At the distance of 21.4 meters from the centre, by the preceding relation of $rv = r'v'$, we get in this case

$$v = \frac{3000}{21.4} = 140 \text{ meters per second}$$

for the gyratory velocity at f, where the pressure is the same as at b.

In the same manner we get for a depression of 500 meters, as at i in the figure, $r = 30.3$ m., and with this, from the relation of $rv = r'v' = 3000$, we get $v = 99$ meters at the distance of this point from the centre, and so at f' at this distance from the centre at the earth's surface, when the pressure is the same as at b'.

The curved lines, Fig. 1, represent the form of the isobaric surfaces where v' is taken equal to 5 meters at the distance of 1000 meters, and where the undisturbed isobaric surfaces AB and $A'B'$ are taken respectively at the heights of 1000 and 500 meters.

235. The gyratory velocity v at the earth's surface, in any isobaric surface, as that represented by $baef$, Fig. 1, is equal to the velocity s acquired by a free body in falling through the height BE of the undisturbed and level isobaric surface, represented by the line AB, and this is given by the expression

$$s^2 = 2gh,$$

in which s is the velocity in meters per second, and h is the height in meters through which the body falls. The demon-

stration of this is too complex to be given here, but it may be found in Recent Advances of Meteorology, p. 241. It is also the theoretical velocity with which a homogeneous fluid issues from a fountain at a level which is at the distance h below the level of the head of the fountain. The value of v, therefore, at f, Fig. 1, is the velocity which a body would acquire in falling through the space BE; at f', that acquired in falling through the distance $B'E$, etc., no account being taken of the effect of friction upon the value of v.

The difference of pressure between f and f' is the same as that between i and f' or that between b and b'; and so the centrifugal force of a horizontal prism or cylinder ff' is exactly equal to the force of gravity on the vertical prism or cylinder if or bb' of the same cross-section.

The value of v is also equal to that acquired by air, at rest under the higher pressure, in passing horizontally to a lower pressure, as for instance of air in passing from E to f, Fig. 1, or to velocity arising in any way from such difference of pressure, and this, when the difference of pressure is not very great, is given approximately by

$$s^2 = 206.3 \frac{P_0}{P} \frac{T}{T_0} \Delta P,$$

in which, as before, s is the velocity in metres per second, P_0 is the standard barometric pressure of 760 mm., P that where the velocity is s, ΔP is the variation of pressure, and T_0 and T have their usual significations, being the absolute temperatures corresponding respectively to P_0 and P. To a given amount of potential energy ΔP corresponds its equivalent in kinetic energy, but the latter, $\frac{1}{2}s^2m$ (§ 18), is proportional to the mass m, and consequently, for a given volume of air, is as the density. But this varies directly as the pressure and inversely as the absolute temperature T, and therefore s^2 must vary inversely as the pressure and directly as the absolute temperature, as the expression above indicates.

By the preceding expression, if the pressure P at f, Fig. 1, is known, and ΔP the difference between f and E, or rather

between that at f and at an infinite distance where $v = 0$, the value of v at f is known, in case ΔP is not very large. Conversely, if v is known, the difference of pressure sensibly between E and f, or rather the whole amount of depression, can be computed. At high altitudes where P is small and the air rare, it is seen from the formula s^2 has to be proportionately large, where ΔP is taken the same at all altitudes.

236. On account of the great depression of the isobaric surfaces, as represented in Fig. 1, the pressure near the centre of tornadoes becomes very much diminished, and in their passage over a place there is sometimes a very sudden change of pressure. Hence, we see the reason of the explosive effects often observed during the passage of tornadoes. Corks fly from empty bottles, cellar doors are burst open against the force of a strong wind blowing against them on the outside, the walls of houses are thrown outward on all sides, the whole roof is suddenly raised up and blown away, or the sudden expansion of air under copper and tin covering rips it up and it is rolled away. And in case the walls which inclose air are not tight, but there are openings through which the air can escape, towels and light articles of clothing which are carried along with the escaping air have been found lodged in them.

"During the tornado of Dec. 22, 1884, in Clarendon County, S. C., a lady, perceiving the approach of a storm, was in the act of closing a glazed door, which extended down to the floor and opened on a piazza; but before she could fasten it the house was enveloped by the tempest, the door flew open, and she was dashed violently against the balustrade running around the piazza, and received injuries and bruises which confined her to bed for several weeks. In the same room there was a heavy pine press, the door of which was locked. This door was burst open, torn from its hinges, and, in the language of the narrator, 'shivered into kindling splinters.' There was no damage done to the house, or at least none mentioned."[66]

In the tornado at St. Cloud, Minnesota, a few years ago, panels were torn from the doors, and with this exception the buildings seem to have been untouched. In other places window panes were blown out and the sash untouched.

The normal air pressure is 14.69 pounds to a square inch.

If, therefore, one-fourth of this pressure is suddenly taken off from the outside of any inclosure containing air of the normal tension, then the effective explosive force is 3.7 pounds to each square inch of the interior surface, or 533 pounds to the square foot. And when we consider the small diameter of a tornado and its rapid progressive motion, it is readily seen how suddenly this explosive force is brought into play, not allowing time for the gradual escape of air and diminution of tension, even where the inclosure of the air is not very tight.

237. In all the preceding relations between gyratory velocity and pressure it has been assumed that there are no frictional resistances between the air and the earth's surface, or between contiguous portions of air with different gyratory velocities at different distances from the centre, and that these velocities follow the law of $rv = c$, as would be the case without friction where the air is drawn in toward, or recedes from, the centre on account of slight centripetal or centrifugal forces. In the case of friction, however, the gyratory velocities can only be maintained by means of a vertical circulation from which arise deflecting forces which overcome the frictional resistances to the gyrations. In the case of friction, therefore, we cannot have a strictly gyratory velocity, as we could if there were no friction to be overcome, but besides the gyratory velocity v, also two other components of velocity, the one the velocity u, of radial motion toward or from the centre, and the other the velocity x, of vertical motion. Putting, therefore, s for the velocity of the resultant motion, we have

$$s^2 = x^2 + u^2 + v^2.$$

But in this case s is less than v in the preceding, where there is the same difference of pressure between the points of initial and final velocity, since in the case of friction a small part of the force due to difference of pressure is spent in overcoming the frictional resistance, and only the balance in producing kinetic energy. So that although we have the additional components of velocity x and u, yet v is so much diminished as to

make the value of the resultant s less than v would be in the case of no resistances with the same differences of pressure.

In the case of friction, therefore, in which it is necessary to have a vertical in connection with a gyratory circulation, the motions in the interior and lower part of the tornado are in toward and around the centre, while the air also ascends, but in the upper part it is around and from the centre, the air still ascending. But the gyratory velocity near the centre of the tornado is usually so great in comparison with that of the radial and vertical motions that x^2 and u^2 do not add much to the value of s^2 in the preceding expression, and so s does not generally differ much from v. The case is similar to that of a right-angled triangle in which one leg is considerably shorter than the other and the square of the smaller one adds but little to the sum of the squares, and so the hypothenuse does not differ much in length from the longer leg.

Since in the case of friction a part of the forces arising from differences of pressure goes toward overcoming the frictional resistances, it is evident that the theoretical relations which have been obtained between velocities and differences of pressure upon the hypothesis of no friction, is not strictly applicable in the usual cases where there is more or less friction, but often considerable allowances must be made for its effect; so that the velocities corresponding to differences of pressure are never quite so great as they would be in the case of no friction; just as, in the case of fountains, the water does not issue from the orifice with that velocity with which it would in case of no friction, and so is not thrown vertically upward to the height of the head of the fountain. The gyratory velocities also, denoted by v, in the case of friction are less in the interior than those given by the law $rv = c$, where c, depending upon the initial gyrations of the air, is known.

THE ENERGY OF A TORNADO.

238. In case of friction, we have seen, a gyratory circulation cannot be maintained without the vertical, nor the latter

without a supply of energy to overcome the frictional resistances to such a circulation. This is found in the supply of heat which originates and maintains the vertical circulation by inducing and keeping up the unstable state; for as long as this state is maintained, the central part of the tornado, where the air ascends, is warmer than the surrounding air, from which arises a force which maintains the vertical circulation. The latent heat of the vapor given out in condensation as the air ascends plays an important part, and is perhaps absolutely essential, since in the case of dry air the unstable state could not be induced up to a considerable altitude, and maintained sufficiently long to give rise to a tornado, as it would require a vertical gradient of temperature decreasing with the increase of altitude at a rate greater than 1° C. for each 100 meters. In this case the heat would have to be continually supplied to the lower strata as fast as it would be diminished by the vertical interchange between the lower and upper strata, in order to keep up the unstable state.

In the case of saturated ascending air, we have seen that the unstable state is maintained with a vertical temperature gradient which is only about half as great, and one which is readily induced in the atmosphere and easily maintained, and so the supply of air nearly or quite saturated for the ascending current is very important, since the latent heat furnished becomes available energy in maintaining the interior part of the tornado warmer than the surrounding part with a much lower temperature in the lower strata, and a smaller vertical gradient of decreasing temperature.

Thermal energy is available in doing work of any kind, or in producing kinetic energy, only where there are differences of temperature, that is, temperature gradients. For instance, if the whole atmosphere were heated up equally in the equatorial and the polar regions to a very high temperature, there would be no general circulation of the atmosphere, and no expenditure of thermal energy in producing motion and kinetic energy and overcoming the resistances to motion. In the case of a tornado it requires not only a vertical temperature gra-

dient, but this must be such as to induce the unstable state. In this case the effective force in maintaining an ascending current and a vertical circulation depends upon the difference of temperature, or temperature gradient, taken along an isobaric surface; so that the greater this difference of temperature between the interior and ascending air and the surrounding air, the greater is this force. But this difference cannot be maintained unless the atmosphere is kept in the unstable state for saturated air by the continual supply of heat to the lower strata as it is drawn away by the ascending current. If, however, the atmosphere is not merely unstable, but considerably beyond the neutral condition, before the ascending current and vertical circulation set in, there is a considerable reserve of energy, which suffices to maintain the motion, until the unstable state is gradually reduced either to the neutral or the stable state. And this latter state can be produced not only by the convection of heat in the ascending air currents, but also by the supply of air for the ascending current which is gradually becoming drier, since the drier the air the greater the difference of temperature required between the lower and the upper strata, and for the lower part, up to where the vapor in the ascending current begins to condense, the gradient of decreasing temperature has to be more than 1° C. for each 100 meters. As there is no expenditure of energy until motion commences, so the more rapid this motion, the greater the rate at which energy is expended.

239. In the origination and first starting of a tornado, the force which overcomes the inertia of the air and causes motion, is the difference of pressure depending upon the difference of temperature between the interior and exterior. This, in the lower strata of the atmosphere, is a centripetal force, and as the air is drawn in toward the centre the velocity is accelerated, and where there is initial gyratory motion, the body is at the same time deflected from its course; but the velocity at any time is the same as if the same force had acted during the same time in a radial direction without any gyratory motion and deflecting force, for the latter, we have seen, does not increase

the kinetic energy, but simply changes the direction of motion. After the vertical and gyratory circulations are fully established, the whole force arising from difference of pressure due to difference of temperature goes to overcome the frictional resistances, and no part to creating motion and kinetic energy, and the velocity of the vertical circulation is such that the resistances are exactly equal to the forces. In the case of no friction, no forces are needed to maintain the circulation after it is once inaugurated, and so no difference of temperature between the interior and the exterior and expenditure of heat energy, but if there is such, it tends to continually accelerate the vertical circulation. The kinetic energy of the gyratory velocity has its exact equivalent in the potential energy of the difference of pressure caused by the centrifugal force of the gyrations, and where there is friction and a necessity for a vertical circulation, there must be a difference of temperature and a corresponding difference of pressure arising from it, to overcome the frictional resistances to this circulation.

In a cyclone, for reasons given in § 232, the forces are mostly spent in overcoming the resistances, while in a tornado they are spent mostly upon the inertia of the air, and so the relations between gyratory velocities and distances from the centre are nearly those given by the law of $rv = c$, and between these velocities and differences of pressure, or pressure gradients, those given in § 233, determined upon the principle that the difference of pressure is caused almost entirely by the centrifugal force, which of course is only approximately so in the case of friction, since a little force and difference of pressure, even in the open air at some distance above the earth's surface, is necessary to maintain the vertical circulation.

240. In a vertical circulation arising from the unstable state, where there is no gyratory motion, the force which gives rise to and maintains the ascending current depends upon the difference of pressure in the column of ascending air and the surrounding undisturbed air at the same levels; and if we assume that the force is spent mostly upon the inertia of the air, and but little in overcoming the friction, we can determine

approximately the velocity of the ascending current from the differences of temperature and pressure, where these are known. At the earth's surface the vertical component of velocity is nothing, and the pressure at the base of the ascending column, if we neglect the friction of the current, is the same as at the earth's surface in the surrounding parts. The differences of temperature in the ascending current and the surrounding undisturbed air, in the case of tornadoes is the same as in the case of cyclones between the columns A and C or C' in the table of § 159, or between D and F, or G and I. Of course these differences depend very much upon the assumed vertical gradient of temperature and the hygrometric state of the air, as exemplified in the table, and there may be cases in nature in which the temperature differences between a column of ascending air and one in the exterior undisturbed air are much greater than those in the table under the assumed conditions. The maximum differences, it is seen, are about 5° C. Let us assume that the average difference of temperature, taken with regard to mass and not volume, in the column of ascending air and one in the surrounding undisturbed air, up to the altitude where the pressure at the same level is equal, is 3° C., and that the level of equal pressure is at the height where the barometric pressure is equal to 380 mm. Supposing the mean temperature of the air column below the level of equal pressure to be 0° C., then the difference of pressure of the column of ascending air below this level and of a similar column in the surrounding undisturbed air, and the effective potential energy in causing kinetic energy at the height of the column, is the 3/273 part of 380 mm., or 4.2 mm. With this value of $\varDelta P$ in the last expression of s^2 in § 235, putting $P = 380$ and $T = T_0$, as it is very nearly by the table of § 13, we get $s = 42$ m., nearly, for the velocity per second due to the difference of pressure between the two columns. Or by Table VI we have the value of h for each millimeter of barometric pressure at the pressure of 380 mm. and at the temperature of 0° C. equal to 21.03 m,. and so for 4.2 mm. equal to 88.33 m., and with this value of h multiplied into $2g = 19.61$ m., we get by the first form of the

expression of s^2 in § 235, the same value of s as above by the last form. This theoretical result is, of course, considerably diminished by the effect of friction, which is not taken into the account in the theory.

241. Where the atmosphere is in the unstable state all the way up to the top, or even to very high altitudes only, the level at which the pressures in the ascending air and at the same level in the surrounding undisturbed air are the same, where there is no depression in the interior from gyratory motion, is very high, and the effective force in causing ascending velocity is great. But if the unstable state of the atmosphere does not extend very high, as is undoubtedly the case often, the horizontal isobaric surface is not very high. The unstable state may extend to so low a level only, that the kinetic energy of the ascending current may be destroyed by the pressure of the column of air before the current reaches the top of the atmosphere. In that case the ascending current is deflected off laterally in all directions below this level, and the isobaric surfaces above that level are not affected directly by the thermal conditions, though they are all, together with those below, depressed by the gyratory motion of the tornado; but as the vertical circulation in this case depending upon differences of temperature extends only up to a certain height, the air above this can only be brought into the gyration by the friction between it and the gyrating air below, and an inverse vertical circulation induced indirectly in the air above. For while the gyrations are more rapid below than above, the tendency is to carry the comparatively little disturbed air above out and around the centre, and to draw down that above in the central part, just as has been explained in the case of cyclones, where the vertical circulation from temperature disturbances does not extend to the top of the atmosphere, and the interior dry and clear air above is brought down into the central part and causes the "eye" of the cyclonic storm. In this way, if the duration of the tornado were long enough, and the atmosphere were calm, or had the same general motion above and below, the gyratory motion would always be induced

from bottom to top, though with less force; but the time of duration being generally short, and the general easterly motion above being mostly much greater than below, tornadoes of this sort, in which the atmosphere above is not in the unstable state, perhaps rarely extend with much violence to very high altitudes in the atmosphere. They may, notwithstanding, have great energy and violence, for the ascending air escapes laterally on all sides above; but of course tornadoes, in general, are of the greatest violence, and the ascending currents most rapid, where there are gyrations and a rarefied centre extending to great altitudes, just as the taller a flue, the greater the draft with the same difference of temperature within and without.

As in cyclones, as may be seen from the examples in the table of § 159, so in tornadoes, the greatest temperature differences are not in the lower strata of the atmosphere, but up at a certain altitude above; or if the atmosphere is in the unstable state all the way up, this difference increases to the top. The conditions may even be such, especially if the air is not nearly saturated, that there is no difference of temperature below between the interior and exterior part, or the interior part up to some height may even be colder. In such cases the energy is mostly at a considerable altitude, and the vertical circulation and the gyrations are first originated there, and gradually communicated to the strata below by friction and by the relieving of the lower strata in the central part of a part of the pressure from the air above, which finally gives rise to an ascent of air from the stratum next the earth's surface and a drawing of it in from all sides at the surface toward the centre.

242. If the gyratory velocity near the earth's surface in a tornado were as great as it is above, there would not be, in general, a very strong ascending current in the interior, for although there might be an enormous difference of pressure, yet the force arising from this difference of pressure with which the air would tend to run in below toward the centre and up in the interior would be in a great measure counteracted by the centrifugal force of the gyratory velocity below. When the

gyratory velocities are the same at all altitudes, the only force by which it flows in below and up the interior is that due to difference of temperature between the central and exterior parts of the isobaric surfaces, by which the vertical circulation is maintained. But even this, we have seen (§ 240), under assumed not improbable conditions, may give rise to a considerable ascensional velocity. This, however, is very greatly increased in consequence of the great frictional resistance to the rapid gyration of the air near the earth's surface, by which the gyratory velocity and centrifugal force here, which tend to prevent the air from rushing into the partial vacuum, are much diminished.

In the assumed example of § 234 the difference of pressure between the points f and E, Fig. 1, is equal to the pressure of the column of air bE, that is, equal to the pressure of a column of air of 1000 meters in height and of the average density between the earth's surface and that altitude, which, in barometric measure at an ordinary medium temperature of 20° C., as may be readily determined by table VI, is about 84 mm., since 1 mm. of barometric pressure corresponds to about 12 meters of altitude. If we suppose the gyratory velocity near the earth's surface beyond f, taking no account of those between f and the centre, to be so diminished by friction that the centrifugal force would be diminished one third, the difference of barometric pressure between f and E, which would be effective in causing an inflow toward the centre, in addition to that considered above due directly to temperature differences, would be 28 mm. This would give rise to a very great ascensional velocity in the interior of the tornado, not however near the earth's surface but at some distance above, for the kinetic energy of the inrush of air from all sides toward the centre would be changed to potential energy in the form of increased pressure in the central part of the tornado at the earth's surface, which would be equal to that at E if the whole centrifugal force at the earth's surface were destroyed, and then this increased pressure would give rise again to kinetic energy in a rapidly ascending current in

the interior, where the air is comparatively rare and the pressure much diminished from the effect of the gyrations above.

With the value of $\Delta P = 28$ mm. or the corresponding value of h taken from Table VI, either of the expressions of s' in § 235, applied as in § 240, gives for a temperature of 20° C., $s = 78.5$ meters per second, or about 176 miles per hour. Combined with this ascensional velocity would be the gyratory velocity of the ascending air, which would be diminished by the effect of friction near the earth's surface, for the ascending air near the centre would consist mostly of the air entering near the earth's surface.

243. The tornado being a column of rapidly gyrating air, of great height generally in comparison with its diameter, and with the air very much rarefied in its interior from the effect of the rapid gyrations near the centre, may be likened to a tall flue with heated and rarefied air in its interior. If the draft of the flue is almost entirely cut off below, the ascending current is feeble, but if the air below has free access, the velocity of the ascending current in the flue becomes very great, and in proportion to the difference of temperature between the interior and exterior. So in a tornado, if the access of air were almost entirely cut off from the lower central part by the centrifugal force of the undiminished gyratory velocities near the earth's surface, the ascending current in the interior would be of no great violence, generally such perhaps as in the assumed cases of § 240; but with a somewhat free access of air below into the rarefied interior, on account of a decrease by friction of the gyratory velocity near the earth's surface, the velocity of the uprush of air in the interior becomes enormous. In this case the centrifugal force of the gyrations in the open air a little above the earth's surface, where there is little friction, almost entirely excludes the access of air into the interior, and serves very nearly the same purpose as the wall of the flue, while the partial destruction of these gyrations near the earth's surface by friction is similar to the letting on the draft of a flue which had been almost entirely cut off.

With an increase of the force of the ascending current, and of kinetic energy, there is a corresponding increase in the expenditure of heat energy, for the more rapid the current the faster is the condensation of aqueous vapor and the freeing of latent heat, and the faster is the heat of the lower strata conveyed to the upper ones by which the unstable state is destroyed unless heat is constantly supplied to the lower strata. The part which friction plays in this matter is the facilitating the application of the heat energy by allowing a more free and rapid vertical circulation and vapor condensation.

FORCE OF THE WIND AND SUPPORTING POWER OF ASCENDING CURRENTS.

244. In view of the enormous gyratory and ascensional velocities in a tornado, it is interesting and important to know the destructive force of the former and the supporting power of the latter. These are both deducible from the first of the expressions of s^2 in § 235. According to this, when air with a velocity of s impinges against a barrier at right angles to the direction of motion, so that velocity in this direction is completely destroyed, or $s_0 = 0$, it gives rise to a force which supports a column of mercury of the height $\varDelta P$, determined by this expression where P and T are known. But if instead of a mercurial barometer we use one of pure water of standard temperature, and denote the change in the height of the column corresponding to $\varDelta P$ by $\varDelta p$, we have $\varDelta p = 13.596\ \varDelta P$, the pressure of mercury being so many times greater than water, and we then get, if we express $\varDelta p$ in centimeters instead of millimeters,

$$s^2 = \frac{2063}{13.596}\frac{P_0}{P}\frac{T}{T_0}\varDelta p = 151.7\frac{P_0}{P}\frac{T}{T_0}\varDelta p.$$

If we now consider the pressure on a unit of surface of one square centimeter, then, since by definition a cubic centimeter of pure water of standard temperature is a *gram*, the pressure of the column of water of the height $\varDelta p$ and unit base in

grams, is Δp, and the pressure upon a surface of which the area is A, is $A\Delta p$, a *gram of pressure* being the gravitational pressure of a gram. Putting p for the pressure of air in motion upon a normal surface area of A, we have $p = A\Delta p$, and with this the preceding expression gives

$$p = .00659 \frac{P}{P_0} \frac{T}{T_0} As^2.$$

The pressure, therefore, is as the square of the velocity and as the density of the air, this latter being directly as the pressure and inversely as the absolute temperature. Hence, at high altitudes where P is much less than P_0, the pressure of the wind for the same velocity is much less than near the surface. This is for pure and dry air. For air with an average amount of moisture in it, it is more accurately expressed by

$$p = \frac{.00659}{1 + .004\tau} \frac{P}{P_0} As^2.$$

The same formula in English measures, in which p is expressed in pounds avoirdupois, s in miles per hour, and A in square feet, is

$$p = \frac{.002698}{1 + .004\tau} \frac{P}{P_0} As^2.$$

With this formula we get for the pressure of wind of the velocity of 80 miles per hour upon a square foot at the earth's surface, where $P = P_0$, when the temperature τ is 25°C., $p =$ 15.7 pounds. At the altitude where $P = 380$ mm., with the same temperature, the pressure for the same velocity would be only half as much. With the temperature equal 0° C., instead of 25° C., the pressure would be increased in the ratio of 10 to 11.

245. The preceding formulæ express the pressure of the wind arising from the momentum of the air, or the change of kinetic into the potential energy of pressure. But this, as in other cases where friction is not taken into account in the theory, is considerably modified by the effect of friction; and in this case the effect is to increase the effect, and not, as is

usually the case, to diminish it. What is usually called *the force of the wind* upon any given object, as a plate of a given area, is the difference of pressure on the two sides. On the one side the pressure is increased, not only by the momentum of the air, but likewise by the dragging effect through friction of the air which passes by, upon the cone or pyramid of comparatively stationary air in front of the plate or barrier. On the other side the effect, to the same amount, is to drag the air away and to diminish the pressure of the air. Hence the theoretical difference of pressure, or effective force of the wind, is increased by equal amounts by the effect of friction upon the pressure of both sides. Accordingly the force of wind determined experimentally is found to be a little greater than the theoretical pressure given by the formulæ.

According to Hagen's empirical formula, determined from very accurate experiments made a few years ago by means of a whirling apparatus, the experimental pressure of the wind on small plates, in grams, denoted here by p', is[66]

$$p' = (0.00707 + 0.0001125u)Fv^2,$$

in which u is the periphery and F the surface of the plate in decimeters, and v is the velocity per second in decimeters. The average barometric pressure in the experiments was 758mm., but no account seems to have been taken of the temperature. If we reduce this expression of p to English measures and the standard pressure, we get

$$p' = (0.00289 + 0.000140u)As^2,$$

in which the notation is the same as in the preceding theoretical expression of p in English measures.

Supposing Hagen's experiments to have been made at a temperature of 15 C., a very comfortable working temperature, and reducing, for comparison, the preceding theoretical expression of p to this temperature, we get

$$p = 0.00255As^2.$$

The difference of these two expressions,

$$p' - p = (0.00034 + 0.0001404u)\,As^2,$$

is the effect of friction in increasing the theoretical value of p.

The plates used were small, ranging from about two to six inches square, and the velocities in the experiments were also small. Supposing the average size of the plates to be five inches square, the value of u become 20 inches, equal to 1.667 feet. With this value of u we get for the experimental expression representing the average for all the velocities and for all the plates

$$p' - p = 0.00057As^2.$$

Hence it is seen that the theoretical coefficient is increased about two-ninths by the effect of friction.

While Hagen's experimental formula may very nearly give the true force of the wind for the average of the plates and velocities used, it is obviously at fault with regard to the law of variation with variations of u and s. By increasing the periphery of the plate, the force, per unit of surface, is increased enormously by his law, and the part $p' - p$, for very large plates, in which u is large and the term depending upon it, becomes the principal term, being very nearly as the cubes of the linear dimensions of the plates, or as uA. For such a law there does not seem to be any reason in theory; and more numerous and accurate experiments, made with plates of a larger range of sizes, would undoubtedly give a different law. It is improbable, therefore, that the formula is applicable to plates of a size much greater than those used in the experiments. Again, the formula makes the part arising from friction to be as the square of the velocity, which is not in accordance with either theory or experiment. The theory of the viscosity of gases would make it as the first power, and hence for high velocities the force given by the formula must be too great even for small plates.

From what precedes, it may be inferred that about one-fourth or one-fifth part must be added to the theoretical

values of wind-pressures per square foot of normal surface given by theory, to obtain the effective force of the wind in moving objects, this force being increased by the diminution of the ordinary pressure of the air on the other side opposite to that on which the wind strikes.

246. The force with which the wind tends to move a spherical body is one-half of that with which it tends to move a body with a normal surface exposed to the wind which is equal to the maximum section of the sphere. The theoretical formula, therefore, which gives the pressure p of the wind against a globe, or the resistance of the air to a globe moving through it, is

$$p = \frac{0.00135 \frac{P}{P_0}}{1 + .004\tau} As^2,$$

in which the notation is the same as in § 243. But here, as in the case of wind-pressure on plates, the theoretical pressure and resistance are increased by friction and for the same reasons.

From the two sets of experiments made at the request of Newton, in St. Paul's Cathedral, at London,—the one with several hollow glass globes of about 5 inches in diameter, let fall from an elevation of 220 feet, and the second one with several bladders formed into spheres about 5 inches in diameter, let fall from a height of 272 feet, the weights being known and the times of descent being carefully noted,—Loomis[44] obtained from the first series, for the coefficient of resistance to a globe 5 inches in diameter, the velocity being in feet per second, 0.0013455, and from the second series, with the bladders, 0.0012693, in ounces Troy. The average is 0.0013074. Hence the expression of the resistance to such a globe, or of the pressure of the wind against such a globe, denoted here by p', is, in ounces Troy,

$$p' = 0.0013074As^2,$$

in which s is in feet per second. The reduction of this formula to the units of measure of § 244, that is, pounds avoirdupois, miles, and hours, gives

$$p' = 0.001421As^2.$$

But in these experiments, as in the case of Hagen's, no account seems to have been taken of the temperature, and so we are at a loss to know what temperature to use in the preceding theoretical expression of p in order to make a comparison of the experimental coefficient with the theoretical. But assuming in this case, as in the other, that the experiments were made at a temperature of 15° C., and also that the barometric pressure was 760 mm., then the preceding theoretical expression of p, for $P = P_0$ and $\tau = 15°$, becomes

$$p = 0.001274As^2.$$

We therefore get

$$p' - p = 0.000147As^2.$$

Hence the theoretical coefficient is increased about one-ninth part by the effect of friction in the case of spheres, according to the preceding experiments.

But Loomis also determined another coefficient of resistance from the results of experiments made by Hutton with a whirling machine and a sphere of pasteboard 6⅝ inches in diameter, for velocities from 3 to 20 feet per second. This was found to be one-fourth greater than the other. This great difference indicates that there is considerable uncertainty in such experiments. Adding one-fourth part to the preceding experimental coefficient, we get in this case

$$p' = 0.001776As^2.$$

Combining this with the previous experimental expression, giving the former one double weight, and supposing all the experiments to have been made under the same pressure and temperature, we get

$$p' = 0.00154As^2.$$

Comparing this with the preceding theoretical expression reduced to the temperature of 15° C., we find that, with this combination of the results of the three series of experiments, the effect of friction increases the theoretical coefficient about two-ninths, the same as in the case of the wind blowing normally against a plate. We may, therefore, assume with considerable probability that the theoretical coefficient 0.002698, in the case of a plate, is increased two-ninths by the effect of friction, and we therefore get for the whole wind pressure against a normal plate

$$p' = \frac{0.00330}{1 + 0.004\tau} \frac{P}{P_0} As^2,$$

and for the pressure against a sphere, or the resistance to a sphere moving through the air with velocity s,

$$p' = \frac{0.00165}{1 + 0.004\tau} \frac{P}{P_0} As^2.$$

From the former of these the values of p' in Table VII have been computed for the several velocities and barometric pressures given as arguments in the table, the temperatures used being those of the table of § 13, corresponding to the pressures at different altitudes. These temperatures may be regarded as a sort of annual means in the middle latitudes, corresponding to the altitudes of the given pressures. For extreme temperatures, deviating considerably from these, slight corrections would be necessary if great accuracy were required.

According to this table and the formula, the mechanical force of the wind in tornadoes, upon surfaces exposed normally to its direction must often be enormous. In the example of § 234, although the gyratory velocity of the air at the distance of 1000 meters is only 3 meters per second, yet by theory the velocity at the distance of 21.4 meters is 140 meters per second, or about 310 miles an hour. This by the preceding formula and the table would give, at the earth's surface, for an average temperature, an effective force in moving an object of about 300 pounds for each square foot of normal surface ex-

posed to the wind. But if the body were a sphere it would be only half as much for each square foot of area of maximum section.

In the case of the velocity of the ascending current of 176 miles per hour near the earth's surface obtained upon the assumptions of § 242, Table VII gives a supporting power of more than 90 pounds to the square foot.

247. Having now determined the force of an ascending current of given velocity s against a sphere of given maximum sectional area A, as in the last expression of p', this becomes the supporting power of an ascending current in keeping such a sphere from falling to the earth. The mass of a sphere of pure water one foot in diameter, expressed in pounds avoirdupois, is $62.43 \times 0.5236\ D^3$, in which D is the diameter of the sphere in feet, 62.43 being the mass of a cubic foot of water, and 0.5236 the ratio of a sphere to its circumscribing cube. Hence the mass M of any sphere of diameter D and density ρ, and likewise the force of gravity expressed in pounds, taking no account of the slight variations of gravity at different latitudes and altitudes, is $M = 32.69\ D^3\ \rho$.

In the case of a body falling through the atmosphere, or any other resisting medium, the velocity of the body is accelerated until the resistance becomes equal to the force of gravity, after which it continues uniform unless the force or the density of the medium changes. But if the ascending current has the same velocity as that with which the body would fall through it after the velocity becomes uniform, the force p' of the ascending current against the sphere, which is the resistance in the case of the falling body, is equal to the force of gravity, and the body is supported. Hence, putting the preceding expression of p' equal to the force of gravity, M, we get

$$\frac{0.00165}{1 + 0.004\tau}\frac{P}{P_0}As^2 = 32.69D^3\rho,$$

as the condition for determining the diameter of a sphere of given density ρ which would be supported by an ascending cur-

rent of velocity s, under any given condition of pressure P and temperature τ.

Since A is the maximum section of the sphere, we have $A = 0.7854D^2$, 0.7854 being the ratio of the circle to its circumscribing square. With this value of A, the preceding equation gives

$$D = \frac{0.0000396}{1 + 0.004\tau} \frac{P}{P_0} \frac{s^2}{\rho}.$$

Having used the expression of p' instead of p, the theoretical expression of the force of the wind, this expression takes into account the effect of friction as heretofore determined.

For the diameter of a sphere of the density of water, which is supported by an ascending current of 100 miles per hour near the earth's surface, where P may be put equal to P_0, assuming that the temperature is 0° C., we get, very nearly, $D = 0.4$ of a foot, or 4.8 inches. For a sphere of ice of density 0.92 at the altitude where $P = \frac{1}{2}P_0$, it gives for the same ascending velocity $D = 0.215$ of a foot, or 2.58 inches.

With the preceding expression the values of D in Table VII have been computed, giving the diameters, in inches, of a sphere of the density of water which would be supported by ascending currents of air of the velocities s in the first column, at the altitudes where the barometric pressures are as given at the heads of the columns, the temperatures being assumed to be as in the table of § 13. For spheres of other densities, it is seen from the formula the tabular numbers must be divided by the density as compared with water.

248. In the case of an ascending current of velocity s, a body with a diameter of the value of D given by the preceding formula, or by Table VII, would remain suspended in the air, but if the ascending velocity were increased, the body would be carried up at such a rate as to leave the relative ascending velocity between the air and the body equal to s. According to the preceding example of a sphere of ice, a hailstone of the diameter of 2.58 inches, up where the pressure in only 380 mm. would be sustained in the air at the same level, by an ascend-

ing current of 100 miles per hour, but if this current were increased to 110 miles, then the hailstone would be carried up at the rate of 10 miles per hour, as long as the pressure and density were not sensibly changed, and it would continue to ascend with decreasing velocity until the value of P in the preceding expression of D would be satisfied for $D = 0.215$ and $s = 100$.

The ascending velocity required to sustain a rain-drop in the air is given by the preceding formula and Table VII. If the rain-drop is 0.1 of an inch, then $D = 0.0083$ of a foot. With this value of D the preceding formula is satisfied up at the altitude where $P = 600$ mm. and $\tau = 10°$ C., with the value of $s = 16.6$, which is a very gentle ascending current, but sufficiently strong to prevent the falling of ordinary rain-drops.

Mr. Dines has measured rain-drops as small as 0.0033 inch in diameter. Drops of this size satisfy the formula up where $P = 600$ mm. with a value of $s = 3$, and hence an ascending velocity of 3 miles per hour would sustain such drops. Mr. Dines[67] has also measured fog particles as small as 0.00062 inch, equal 0.0000517 of a foot. With this value of D the formula at an altitude where $P = 600$ mm. gives $s = 0.41$. The law of resistance as the square of the diameter and velocity, as given by the expression of p', § 246 (for A is as D^2) may not hold very accurately for so small particles, but the formula applied in these cases at least shows that exceedingly small ascending velocities are sufficient to sustain even the larger and measurable cloud and fog particles in the air, and so there is no necessity to resort to the old and improbable vesicular theory of cloud particles in order to account for the floating of clouds in the air. For in the cyclone and cloud areas we know there are always ascending currents, and in the surrounding parts where there is no ascent of the air, or generally a slight descent, there is a clearing off from the descent of the cloud particles into the unsaturated air below, where they are evaporated and converted back again into clear and transparent vapor. Hence, as in the vertical circulations of the atmosphere the slight ascending currents in some places give

rise, first to haziness, and then to a clouded sky, so at others, where there are very slowly descending currents, the sky which previously was entirely overcast with clouds becomes, somewhat all at once, clear from the vanishing, and not from the passing over of the clouds.

The size of rain-drops in all cases is an indication, in some measure, of the velocity of the ascending currents. Small rain-drops indicate that the ascending current is so feeble that they can fall, and are not carried up; but where only very large drops fall, the indication is that the ascending current is so strong that all except the very large ones are either carried upward, and in the case of a tornado outward above to where the ascending currents are less, or else are sustained in the atmosphere until by the contact and blending with them of other drops, they become large enough to fall.

249. Only a few of the many instances of observed force of the wind and supporting power of ascending currents in tornadoes can be given here, for the literature on this subject has become immense. From the account of the tornado of April 16, 1875, at Walterborough, S. C.,[69] the following facts are taken:

The Academy and seven churches were totally destroyed, and of the whole number of ninety dwellings some sixty were torn to pieces and strewn to the four winds of heaven. At least 5000 trees in the village were scattered in every direction. The roar of the wind was such that the tremendous crash was heard by no one. One gentleman was looking from his piazza and saw large trees fall in his yard, but heard no more sound than if they had been feathers.

The greatest destruction was on the south side of the track. A heavy piece of timber 6 by 6 inches and 40 feet long, weighing 600 pounds, was carried from the Episcopal church in a direct line over the tops of houses to a distance of 440 yards in a direction W. 10° N. The wind continued in its first direction for a few seconds only, when it changed to the north with increased fury. A large 4-horse lumber wagon, weighing 3,500

lbs., was lifted over a 6-foot fence and carried 60 feet from its original position.

A chicken coop, box 4 by 4 feet, weight 75 lbs., was carried 4 miles. Hickory trees, 54 inches in circumference at butt, weight 3000 pounds, were lifted out of the ground and carried up a bank 10 feet high. A cart weighing 600 pounds was carried up in the whirl, torn to pieces, and the tire of one wheel found 1320 yards distant. Dead sheep were found with wool off the hide. Geese plucked of feathers, as if picked by hand.

From the account of the tornado of Aug. 9, 1878, at Wallingford, Conn.,[10] the following statements are gleaned: The great power of the wind at one place was illustrated by the carrying of a large oak tree so far that its place of growth could not be found. Persons witnessing it said that two currents of air seemed to unite at this point, where the valley grew narrower.

The indraught reaching down Old Colony street was severe enough to take large elm trees and wrench them off at some distance from the ground. Along this street where the roots of trees were strong and well planted, the limbs only were affected, the appearance being that the force was exerted some feet above the surface. A barn was carried bodily away and no parts of it afterward recognized. Mr. Vasseur's house was lifted up and the parlor floor arched upward. An elm tree, measuring 9 feet in circumference, was broken off 9 or 10 feet from the ground.

Pieces of timber and scantling were imbedded in apple trees so far (fully 6 or 7 inches deep), that in efforts to pull them out they were broken off. An iron rowboat, said to weigh about 80 or 85 pounds, was lifted from the water of the pond and carried by the force of the wind 225 feet. A large apple tree, weighing nearly 1500 pounds, was carried 20 feet due north.

The arching up of Mr. Vasseur's parlor floor was, no doubt, caused by the expansion of the air beneath, from the sudden removal of the surrounding pressure (§ 236).

250. The following facts are taken from the account of the

tornado at Mount Carmel, Illinois, June 4, 1877:" The houses on the south side of the street were totally destroyed, and bricks and other objects hurled with great force northward against the buildings, somewhat damaged but left standing, on the opposite side of the street. A brick moving at an angle of 15° or 20° with the horizon, entered the Harris house through the weather-boarding, lath and plastering, crossed two rooms, a distance of 27 feet, and lodged in a rear wall without breaking even the corners from the brick. From the large and circular aperture through which the brick entered it was thought that it must have whirled rapidly in its transit. So great was its velocity, that the laths were cut quite smoothly without cracking the adjoining plastering.

Lewis Gott saw the tornado strike his house, his point of observation being one square north. As he described it, the house appeared to go up bodily and plunge into the cloud. Only a very small portion of it was ever seen afterward to be recognized.

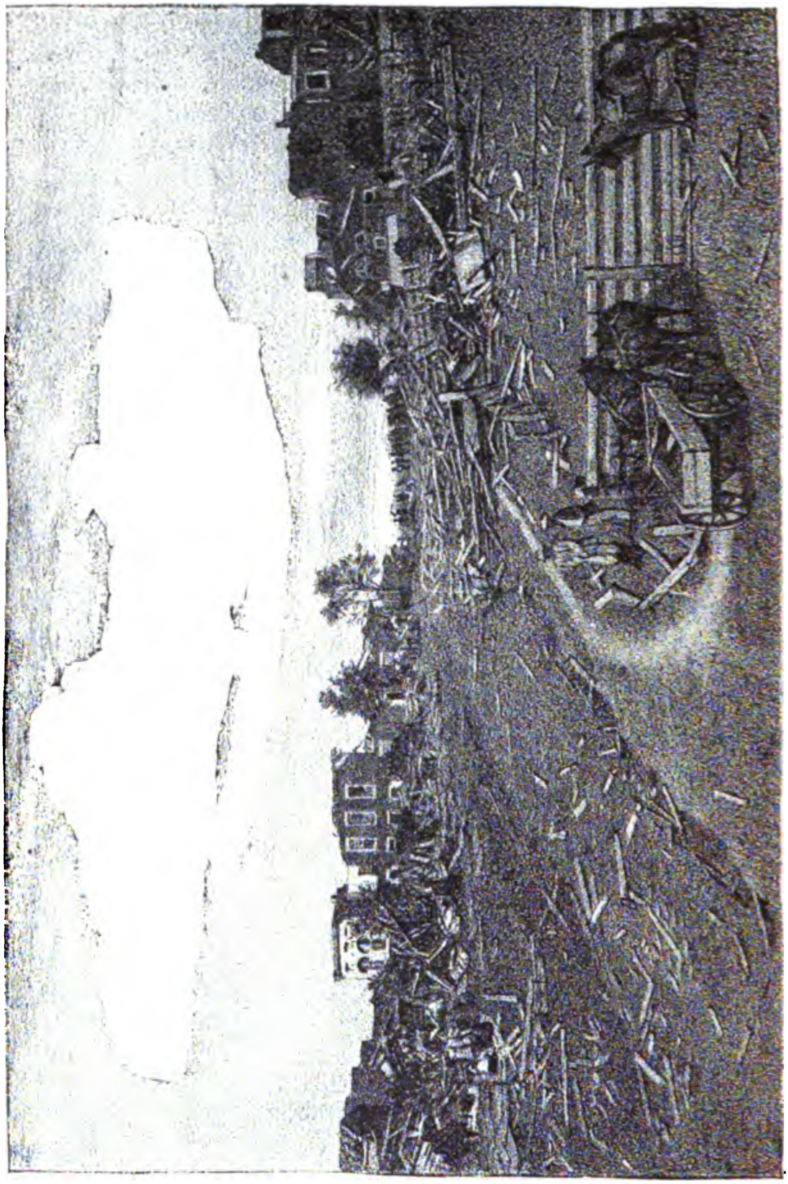
Although many small objects were thrown outward by the whirl of the tornado, yet the general tendency of trees and prostrate buildings was inward toward the centre of the track, this effect being more noticeable, however, on its southern than on its northern side.

Objects were carried to great distances. A piece of tin roof was carried 17 miles N.E. The spire, vane, and gilded ball of the Methodist Church were found at the distance of 15 miles N.E. A letter was carried by the wind 45 miles in the direction of N.N.E. A paper sack of flour was found nearly 5 miles distant in Indiana.

The accompanying plate, copied from the Report of the Chief Signal Officer for 1877, shows the destructive violence of this tornado in a narrow path where it passed through Mount Carmel.

251. The following extracts are taken from an account by H. C. Hovey of the tornado at St. Cloud and Sauk Rapids, Minn., which took place on April 14, 1886."

"During the day a remarkably high temperature, for the season, had



TRACK OF MT. CARMEL TORNADO, FOURTH STREET, LOOKING EAST FROM CHESTNUT STREET.

prevailed, the mercury rising as high as 80°, and the air was sultry and oppressive. At 3 P.M. observers saw dark banks of struggling clouds overhanging the ridge that in ancient times used to be the river limit, and there were apprehensions of impending danger. Suddenly the clouds began to revolve, while sharp points shot downward, until a whirling funnel-shaped mass was formed above a basin amid the hills, that seems to have furnished the cradle for the ensuing tornado. Its first condition was undoubtedly that of a simple whirlwind, having a diameter of about 1000 feet, which uprooted or twisted off nearly every tree in its circle, overturned the monuments in the adjoining Masonic cemetery, and tore up the boulders from the ground. Thence it moved slowly and majestically along at the rate of about 12 or 15 miles per hour, but with an inconceivably rapid rotary motion upon its vertical axis, confining itself for some distance to a path hardly more than 150 feet wide. Hundreds of people took timely warning and got out of the road of the moving column of cloud, whose general trend was toward the northeast. Having wrecked the Catholic Church on Calvary Hill, and also several farm-houses, it entered a portion of the city of St. Cloud mainly occupied by foreigners, whose frame cottages were strewn over the plain indiscriminately, leaving nothing but the cellars to mark the sites of the houses.

"I noticed but one exception to this general work of complete demolition, and that was a house that had been whirled about end for end and left on its foundation as a wreck. Reaching the freight depot of the Manitoba R. R. the wind tore to pieces, and overturned the long line of freight cars, carried the trucks away, and even in places wrenched the iron rails from the ties.

"The tornado struck the Mississippi River at a point opposite the village of Sauk Rapids, and fishermen who were in full view of the crossing aver that for a few moments the bed of the river was swept dry; and in corroboration of this remarkable statement they showed me a marshy spot where no water had been before this event took place. Two spans were torn away from the substantial wagon bridge below the rapids, one span being hurled up stream and the other down it by the rotary motion of the blast; and great blocks of granite being also torn bodily out from the piers. The large flour mill near the bridge was levelled. The depot of the Northern Pacific R. R. was demolished, and the central portion of the village itself was attacked with the greatest violence. Being the county seat, the court-house was located here, a substantial structure, of which only the vault, six iron safes, and the calaboose were left—the latter turned upside down. A fine new school-house, costing $15,000, was completely swept away. The Episcopal Church was so utterly removed that the sole relic thus far found is a battered communion plate. The floor of the skating-rink is all that is left of that structure. Stores,

hotels, a brewery, and four-fifths of the residences in the village were scattered as rubbish along the hillsides, or borne away for miles through the air.

"One of the saddest of the tragedies marking this wide disaster took place at a farm-house in the country, about 16 miles north of Sauk Rapids, where a wedding party of 30 persons were assembled. The ceremony was just concluded, and the officiating clergyman was offering prayer when the building was struck by the tornado. The bridegroom was killed outright, as were also 15 others; 7 more victims have since died, and only one of the company escaped severe injury of some kind."

The number of killed at Sauk Rapids was 39, and about 100 injured more or less.

252. On the 29th and 30th of May, 1879, there were many very destructive tornadoes in the States of Kansas, Nebraska, Iowa, and Missouri, which were thoroughly investigated by Sergeant (now Lieutenant) F. P. Finley of the Signal Service, and the results given in a published report.[n] From this lengthy report of 116 quarto pages, containing a great number and variety of very important and astonishing facts in relation to tornadoes, the following few are selected as examples of the great force of the wind and its raising and supporting power in a tornado, and in corroboration of the theoretical deductions in preceding pages:

From the account of the Lee Summit tornado on the western border of Missouri the following extracts are taken:

"Among several of the peculiar manifestations of the wind's force at this place, I relate the following: 150 feet N.E. of the house and 10 feet W. from the stable, a large gate was carried 200 feet to the S.E. and torn to pieces without injuring the stable. A heavy lumber wagon, standing behind the corn-crib and at a distance of 155 feet east of the house, was lifted up bodily and carried to the S.E. over a corn-field a distance of 100 feet, without injury. Two window-panes were blown from a sash on the W. side of the house and carried inside without breaking them, while two out of the N. window were found broken. A large well-curb lying E. of the house 15 feet was carried one-fourth mile to the S.E. and lodged in a wheat-field. A heavy sulky cultivator, weighing about 600 pounds, was carried free from the ground a distance of 86 yards to the S.S.E. and broken by the fall. Another one standing near the corn-crib was partly carried and partly rolled along a distance of about 60 rods to the S.E., and

then literally twisted to pieces and the *débris* scattered about over a field of 12 acres, apparently in circles."

"The following will indicate some of the peculiar freaks of the storm: A carpet upon the floor of the log part of the house, and securely tacked about the edges, was taken up and carried out of the house without being torn. A new sewing-machine was broken into forty or fifty pieces. Five feather beds were torn into strips and the feathers scattered broadcast over the country. Several garments were carried four or five miles to the N.E. An iron kettle, holding 15 gallons, was found broken into six pieces and scattered about in several directions. A 10-gallon keg filled with vinegar was carried to the N.E. 40 rods. A large iron-bound trunk, fitted with an extra heavy lock, was torn to pieces, and the lock found a half mile to the N.E. sticking into a rail. Several photographs which were known to have been securely placed in an album in the trunk, were found over four miles to the N.E. A watch in the vest of Mr. J. H. Warden, was blown 50 yards to the N.E. and found covered with mud, but the vest was carried to the E. 20 yards. Several chickens were carried to the N.E., from one-fourth to one-half mile and entirely stripped of feathers. Two stoves were broken into small pieces, and one standing near the middle of the house was uninjured. Heavy bed-quilts were so filled with mud that when dry they were as stiff and hard as boards. A lumber wagon was carried to the N.W. 10 rods, the box torn to pieces, and nearly all the spokes taken out of the wheels. An iron-beam plow, 25 feet W. of the house, was not moved or injured, and a seed-drill and harrow near the barn were also untouched."

According to Mr. Richardson's account, at Mr. Hutchins's, "trees and broken timber could be seen, jerked up into the vortex, whirled around with terrible fury, and then thrown outward at the top."

253. The following instances of the great power and fury of a tornado are taken from the account of the Irving tornado:"

"The houses of Edward Wentworth, Gavin Reed, Robert Reed, and Wesley Cooper, situated in the bottom of a 'draw' leading into Game Fork Creek from Timber Creek, in an E.N.E. direction, were next reached and fairly ground to pieces. Hardly a board 6 feet long was left near the foundation of any of the buildings. The funnel as it passed the high 'divide' to the 'draw' perceptibly widened at the bottom; but without bodily swooping downward at once, it drew the buildings up into its vortex and then twisted them to pieces. The house of Robert Reed, 16 by 24 feet, and one and a half stories high, was lifted up as easily as a feather, and without at first cracking the timber. So quickly

was it done, that before Mr. Reed, who was within, knew his danger, the building had risen a height of 25 feet or more. The house being then enveloped in darkness, and not knowing what had happened, he started for the door, thinking it time to make good his escape, when, instead of stepping out upon the ground, as he expected, he fell the above distance, injuring himself severely."

After giving various other details with regard to this tornado, it is further stated:

"Going back to the funnel cloud where we left it at its entrance upon Game Fork Creek, we find that it followed the bed of the stream very closely, hugging the eastern bluffs with such tenacity that it ripped up nearly every tree along their sides and withered the tough prairie grass. Persons who watched its progress along this portion of its track stated that the demoniac fury of the cloud was appalling; whirling with most frightful rapidity, the intense black column would at times seem to level the whole bluff as it disappeared from view within the huge rolling mass of darkness. The eastern bank, covered with a luxuriant growth of timber, would, as the cloud moved along, successively emerge from its awful baptism swept clean to the soil. While this terrific manifestation of force was going on along the stream, westward over the valley, a distance of 60 rods, only a gentle wind was experienced."

254. Of the Stockdale tornado, when it struck Mr. Condray's house, it was said

"that the roof was carried to a perpendicular height of nearly 200 feet. . . . . The roof was lifted as easily as if it had been a feather, shot upward in the very centre of the cloud, and fairly ground to pieces, the *débris* whirling in circles about the upper portion of the cloud, and finally dropping to the ground one-fourth of a mile from the centre of the storm's path."

In the Delphos tornado,[12]

"two new lumber wagons were carried, one 35 and the other 40 rods to the N.W., and broken into pieces; one of the wheels, nearly whole, was found a distance of one mile N.W. of the barn; the other wheels were broken from the axles, divested of their spokes, and the large heavy iron tires were bent into every shape. A log from the barn, 12 feet long and 10 inches in diameter, was carried 320 yards to the N.N.W. The front iron axle of a top buggy ($1\frac{1}{2}$ inches in diameter) was found bent double, the two ends crossing each other, and both wheels were torn off even to the hubs."

It is also stated of the same tornado that heavy cast-iron

wheels broken from a large harvester, and weighing 200 pounds each, were carried half a mile.

In the Barnard tornado

"a large government wagon-box, covered with about 200 pounds of iron in the way of fastenings, was carried away from near the corn-house, and no portion of it afterward discovered. Two sulky cultivators, weighing between 400 and 500 pounds each, were broken into pieces and portions carried to the E. for a distance of one-fourth of a mile."

The Gentry County tornado, in crossing the river at Greenwell ford, 4 miles S. of Albany,

"lifted the new iron bridge of 160 feet span and weighing 120 tons, from the stone pier upon which it rested, carrying it into the bed of the river without overturning it. The structure was strongly bolted to the abutment at one end, while the other end rested on rollers at the opposite abutment to allow for the expansion and contraction of the iron stringers."

255. The prostrating force of horizontal, and the raising and supporting power of ascending, currents, as manifested in the preceding extracts from the accounts of the observed phenomena of tornadoes, are mostly explicable from the theoretical deductions of the preceding pages. By the law of $rv = c$, § 232, we have seen that where there is a centripetal force and an initial gyratory motion, even very small at a considerable distance from the centre, the velocity becomes enormously great near the centre, and there is a great concentration of the energy there. The assumed example of § 234, in which we get a theoretical velocity of 140 meters per second at the distance of 21 meters from the centre, is not an improbable one, which can rarely occur in nature, but there are probably cases which, after making all due allowance for friction, may give much greater velocities at that distance, and these are enormously increased still nearer the centre. But with this velocity we have seen, § 246, the force exerted upon each square foot of normal surface is about 300 pounds. The force of such a velocity, therefore, on the side of a large building, although it were of great weight, is amply sufficient to overthrow it and carry it along until it is broken to pieces and the parts scattered in many directions.

Such a force also is sufficient to move cars from the track of a railroad, of which we have several instances in the preceding extracts, and carry them to a considerable distance.

According to the preceding law, it is seen how the kinetic energy of a tornado may be enormously great in the vortex of a tornado, while at a very short distance there is scarcely any perceptible wind. Thus in the case of the Irving tornado (§ 253), at a distance of only 60 rods from the vortex and place of greatest devastation, only a gentle wind was experienced. Hence the track of destructive violence of a tornado is always narrow. According to Finley,[8]

"The width of the path of destruction, supposed to measure the distance between the areas of sensible winds on the sides of the storm's centre, varied from 40 to 10,000 feet, the average being 1085 feet."

A wind with a velocity of 140 meters per second would of course have much less force against a tree, even with foliage, than against a solid barrier, since much of the air would pass through the tree, and so would exert no force against it except by friction upon the air which is obstructed and does not pass. But making an allowance of one-half, we yet have an enormous force against the top part of the tree, which acts with a leverage upon the stem in breaking it, and upon the roots of the tree in overturning it.

But we have reason to think that the velocity of the air in a tornado, especially very near its centre, is often much more than 140 meters per second, and so by Table VII, the forces exerted against the obstructions are still much greater.

256. From the theoretical result of the example in § 242 of an ascending velocity of 176 miles per hour, which is based upon no unreasonable assumptions, but upon such as can by no means be regarded as extreme ones, we have reason to think that the velocity of the ascending current in the central part of a tornado is often enormously great. Even the velocity of 176 miles per hour, gives a lifting and supporting force near the earth's surface of more than 90 pounds to a square foot (§ 246). Extreme cases may even give forces several times greater than this.

The nearly horizontal and gyratory currents at a little distance from the centre near the earth's surface, where they are more radial than at a small altitude above the earth's surface, necessarily become largely ascending currents also as they approach the vortex of the tornado, and so, with the enormous forces of so great velocities, readily carry large and heavy bodies into the vortex and up in the interior of the tornado, and support them at a considerable altitude above the earth's surface, until, in the progressive motion of the tornado, they have been carried to a great distance. As the altitude is increased and the density of the air diminished, the formula and Table VII show that the supporting power of the same velocity is diminished, and in proportion to the density. It may happen, therefore, that the raising and supporting force of the tornado may be such as to raise a body and support it at a considerable altitude, but not be able to raise it up to where the currents above are outward from the centre, and so it has to remain at some distance above the earth's surface, within the vortex of the tornado, where it is at the same time whirled rapidly round and round the centre, until, from some increase of the energy of the tornado, it is carried up and out above, where the ascending current is not strong enough to prevent its falling, or until for some reason the energy of the tornado is so exhausted, and the velocity of the ascending current so diminished, that it falls directly back in the interior and central part. In either case the tornado may have progressed meanwhile to a long distance. Smaller and lighter bodies are of course carried up at once to great altitudes, if the ascending currents extend up so high, but these may be retained for a considerable time in the central region of the tornado above, where the motion is much more gyratory than radial, before they get so far from the centre that the ascending currents do not keep them from dropping down; for the lighter bodies have to be carried out to a much greater distance from the centre before they can fall.

257. There is not much difficulty, therefore, in accounting for the sustaining in the air and the transportation of light,

and even heavy, bodies to a great distance. Take, for instance, the heavy piece of timber 6 by 6 inches and 40 feet long, weighing 600 pounds, which was carried from the Episcopal Church over the tops of houses, § 249. While it was horizontal it offered a surface of 20 square feet to the wind. By Table VII an ascending velocity of 100 miles per hour would raise and support a weight of 600 pounds with a normal surface of 20 square feet exposed to the current, and this may be considered as no unusual velocity of the ascending air in a tornado. The same velocity perhaps was sufficient to carry up the cart weighing 600 pounds, as the amount of surface exposed to the current was probably at least 20 square feet. Also a much smaller velocity would raise and sustain the chicken coop 4 by 4 feet, weight 75 pounds, while being carried 4 miles.

No unusual velocity, likewise, was necessary to raise and keep up in the air the tin roof at Mt. Carmel, § 250, while it was carried 17 miles N.E., or the spire, vane and gilded ball of the Methodist Church while being carried 15 miles N.E., or a paper sack of flour while being transported 5 miles.

In the same manner the lifting and carrying away of the heavy lumber wagon at Lee's Summit, § 252, or the well-curb, or the cultivator weighing 600 lbs., may be accounted for, for we can assume as probable an ascending velocity of much more than 100 miles per hour, if necessary.

The iron bridge weighing 120 tons, which was lifted by the Gentry County tornado, § 254, although of great weight, had also a great amount of surface exposed to the ascending current. If we suppose that there was a weight on the average of 300 pounds to the square foot, then according to computations § 247, and to Table VII, an ascending velocity of about 320 miles per hour would be necessary to lift it. This is perhaps an extremely great velocity, but the weight of the bridge may have been much less than that assumed above, and so the ascending velocity required much less.

258. We now see how a body may be raised up and sustained in the atmosphere when once up where there are rap-

idly ascending currents, as in the case of the iron bridge above, but it remains to explain how bodies close on the earth's surface, where there cannot be such currents, are raised bodily and vertically up into the air. Several examples of this sort are given in the preceding extracts. The house of Lewis Gott appeared to go up bodily and plunge into the cloud (§ 250), and in the Lincoln County tornado, Mr. Moore's house, 200 yards west of the storm's centre, 12 by 30 feet and one and a half stories high, "was lifted bodily and carried, with the entire family in it, to the centre of the track, in a direction a little S. of E., where it went to pieces."[72] The house also of Robert Reed and of others was lifted up as easily as a feather (§ 253), and likewise the roof of Mr. Condray's house (§ 254), in the Stockdale tornado. In the account of the Irving tornado it is also stated,[73] that

"the house of Mr. Morgan stood nearly in the storm's centre. The wind first struck it on the E. side, lifting it bodily up to the W., off the foundation, when it disappeared out of sight in the dark boiling mass of whirling clouds. Mr. Morgan and family were in the cellar and saw the house pass from over their heads in the manner above described."

In the rapid gyrations of the vortex of a tornado, where there are numerous inequalities of the surface and substantial building obstructions, the horizontal currents may be abruptly deflected upward so as to give a strong vertical current very near the earth's surface, by which objects may receive an upward start, but there are cases in which buildings seem to be drawn vertically upward into the vortex of the tornado from a level surface where there are no inequalities and obstructions of this sort.

Let us suppose that a tornado, by its rapid gyratory motion in the open air above the earth's surface and down to very near the surface, brings down the horizontal and disturbed isobaric surface *AB* in the figure following to the earth's surface, *DE*, or nearly, in the form of the curved lines *ae* and *bf*, as explained in § 233, but that near the earth's surface in the stratum of air between the points *e* and *f* and the surface *DE* there is little or no gyratory motion. This would be very much less

here than above, on account of the friction of the earth's surface, if even the tornado were stationary, but in its progressive motion, carried along by the general progressive currents above, the rapid whirl of the air above does not have time to communicate such gyratory motion to the lower stratum of air, which has less progressive motion, for the vortex above is being continually brought over new portions of the stratum below.

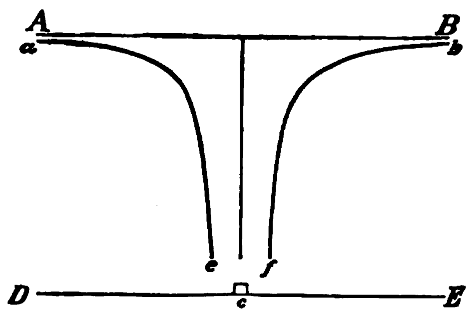


Fig. 2.

Between the points e and f, say at the height of 150 feet above the earth's surface, there may be nearly a vacuum, while at the earth's surface, at c, the pressure and density are very nearly the same as at D and E, since there is, at least at first, scarcely any gyratory motion to counteract the horizontal communication of the pressure from D and E to c. Let us suppose that instead of a vacuum the barometric pressure between e, and f is 15 inches, while at c it is 30 inches. There is then a decrease of barometric pressure between c and the central part of the vortex at the level of e and f of 15 inches, in 150 feet, or one inch for each ten feet, supposing the rate of decrease to be the same through the whole vertical space of 150 feet. If now there is a cubical body, of no very great weight or solidity, as a wooden house, with a linear dimension of 20 feet, at the point c under the vortex, and resting on or being very near the earth's surface, then the height of this body being 20 feet, the upward pressure at the bottom is nearly 15 pounds to the square inch, being the ordinary air pressure at the earth's surface, while at the top, since the barometric pressure is decreased two inches, or one

fifteenth of the whole at the earth's surface, the pressure to the square inch is only 14 pounds. Hence the upward pressure on the lower surface is one pound to the square inch greater than the downward pressure on the upper surface, and the force which tends to raise the whole body vertically upward into the vortex at the level of *e* and *f*, is one pound to the square inch, or nearly 30 tons for the whole body with a superficial area at the bottom and top of 400 square feet. This would be much more than sufficient to raise vertically upward from the earth's surface any wooden structure of those dimensions, and hence, from this assumed example and estimated force, it is seen how bodies upon and close to the earth's surface, where there can be little, if any, ascending velocity of the air, get a start upward. When once a little distance up, where the velocity of the ascending current in consequence of this rapid decrease of pressure is very great, bodies are raised and sustained by the force of the current, as already explained.

RESULTANTS OF GYRATORY AND PROGRESSIVE MOTIONS.

259. According to Finley,[3] of the 600 tornadoes upon which he reported, "the rotary movement of the whirling cloud was invariably from right to left, or the opposite movement of the hands of a watch." This indicates either that the earth's rotation on its axis, as in cyclones, must determine the direction, or that the atmosphere has numerous whirls in this direction, the results of previous cyclonic gyrations which have not been brought entirely to rest. There is little doubt, however, that the direction of gyratory motion is almost always, if not entirely so, contrary to that of the hands of a watch, though reports are not entirely wanting of their rotation the other way.

The direction of the progressive motion of tornadoes is mostly northeasterly.

"Of the tornadoes the courses of which have been recorded, 310 moved from S.W. to N.E.; 38 from N.W. to S.E.; 16 from W.S.W. to E.N.E.; 14 from W. to E.; 7 from S.S.W. to N.N.E.; 5 from W.N.W. to E.S.E.; 3 from N.N.W. to S.S.E." "The velocity of progression of

the storm cloud, as determined from the reports in 130 cases, varied from 12 to 60 miles, the average being 30.08 miles." [18]

The direction of the general drift of the air is very nearly that of the progressive motion of the tornado, and so mostly from S.W. to N.E. The velocity of this is always considerable in comparison with, though generally much less than, the gyratory velocity of the violent part of the tornado. On the right hand, and mostly the southeast side, of the path of the tornado, therefore, the two motions very nearly coincide in direction, and so the velocity of the resultant of the two motions combined becomes much greater than on the left-hand side where the progressive and gyratory motions are somewhat in contrary directions, and where, consequently, the velocity of the resultant motion is very nearly the difference of the two velocities. For this reason the most violent and destructive part of the tornado is found on the right-hand side of the central path, and the tornado, consequently, as the cyclone (§ 202), has its *dangerous side*. Consequently where a tornado passes through a forest the fallen trees are found mostly on the right hand side of the central track, and with their tops in a northeasterly or north-northeasterly direction. The gyratory velocity, however, may be so great that the weaker trees at least are overthrown on the other side, and then they are there found lying mostly in a somewhat contrary direction.

260. On account of the general progressive motion of the air, nearly in the direction of the motion of the tornado, the absolute motion of the air in the front of the tornado is nearly at right angles to the direction of the path, or perhaps generally inclines a little forward, the progressive motion counteracting, and even move, the inclining of the wind toward the centre. But in the rear of the tornado, the progressive motion increases the inclination toward the centre, and here trees are overthrown and objects carried nearly toward the centre and in the direction of the storm's path. The effect upon the resultant direction is similar to that in cyclones, as explained in § 201 and represented in Fig. 9. Where the centre of a tornado passes over a place in a northeasterly direction the trees

in front may be thrown toward the N. or N.W., but in the rear toward the E., or still more around toward the N.E. in the direction of the storm's progress. In consequence of the progressive motion in the rear of a tornado being more nearly in the direction of the gyratory motion, the winds in the rear are usually stronger than in the front part, and so the stronger trees, which are not overthrown in the former, may be prostrated in the latter.

Where the right-hand side of a tornado passes over a place the trees may be first thrown toward the N.W. or N., but mostly by the south quadrant of the tornado toward the N.E., for here the gyratory and progressive motions most nearly coincide in directions, and consequently the resultant velocities are the greatest: and this is the most dangerous part of a storm. But if the left-hand side pass over a place, the trees may be thrown successively toward the N.W., W., and S.W., and by the western quadrant even toward the S. or S.E. Hence it is stated with regard to the Gentry County tornado:[72]

"The largest trees were twisted off near the ground, uprooted, or broken off, and lay with their tops pointing in a circular direction from right to left within the central part of greatest destruction. Those on the S. (right-hand) side of the centre were pointing to the E. and N.E., and even N.W. when very near the centre. On the N. side they were pointing N.W., W., S.W., and S.E."

261. In the progression of a tornado in a northeasterly direction, the trees first thrown down being those of the front part, are thrown more from an easterly direction than those afterward prostrated by the rear part, which are decidedly from a westerly direction, because here the progressive and gyratory motions are somewhat in the same direction. Those from a more easterly direction, therefore, where they cross, generally lie under those of a more westerly direction. This accords with the observations of Dr. Anderson,[68] who says:

"The first area examined, tornado of April 23, 1883, was composed of two distinct parts. The first was a long rectangular space of about half a mile in length, from west-southwest to east-northeast, and a hundred and fifty to two hundred yards in width. Within this space the trees were prostrated from southeast, south, southwest, and west, and inter-

mediate points; and wherever two were found lying across each other, the one thrown from the direction the nearest to east, or farthest around from west, was always at the bottom. Thus, those thrown from south always lay on top of those from southeast, those from west were always on top of all other directions. This order was without exception. The rectangular area terminated at the east end in an irregularly circular area of about eight hundred yards diameter, either east and west or north and south. Bisecting this area both ways and dividing it into four quadrants, the southwest and southeast were found to correspond in all respects with the rectangular area, except that in the southeast there was a greater proportion of trees thrown down from east-southeast and southeast than in the other sections; and in the southwest quadrant, near the centre, a tree thrown from the southwest was overlaid by one from south, the single exception to the order noted above. In the northeast quadrant the destruction was less than in either of the others, and trees were thrown down from east, northeast, north, northwest, and west. In the northwest quadrant the trees were thrown from north, northwest, and west, chiefly from northwest, west-northwest, and west; and in the instances where they crossed each other, the order in relation to the west was similar precisely to that of the other parts, progressing from east round by north to west, as, on the other side, the progression was from east round by south to west; so that in these, the northeast and northwest quadrants, trees thrown from northeast lay under those from north, those from north under those from northwest, while, as in the south quadrants and the rectangular space, those from west were on top of all."

The destruction and prostration of trees in the rectangular area referred to was caused by the dangerous side while the tornado had a progressive motion, the other side not having sufficient violence to prostrate the trees. But at the end of this area the tornado seems to have remained nearly stationary for a short time, when the violence was nearly the same on both sides, and during this time trees were prostrated on the north side, in the northeast and northwest quadrants. The order of prostration in the rectangular strip, and in the southeast and southwest quadrants while the tornado was stationary, was the same from east around by south to the southwest, while in the northeast and northwest quadrants, while the tornado was stationary, the order was the reverse, from the east round by north to west.

On account of the right-hand side of a tornado, as already explained, being the more violent and destructive side, there is generally a marked difference in the width of the destructive parts of the two sides. With regard to this it is stated in the account of the Lees Summit tornado:[72]

"Mr. Bradley and his son, who had been engaged for several days in putting up fences over the path of the storm, were struck with the marked contrast in the comparative width of the E. and W. sides of the storm's track. For a distance of nearly four miles they had found that the W. (left-hand) side averaged about 40 rods in width, the maximum being 80 and the minimum 25 rods. On the E. (right-hand) side, over the same distance, the average width was over a mile, the maximum being two miles and the minimum three fourths of a mile."

Finley also states with regard to the tornado tracks generally:[72]

"The left or W. side was always the narrowest, and, unless within the limits of the path of greatest destruction, the force manifested was always the weakest on this side. I inquired repeatedly if any one experienced strong inrushing currents on this side which were often mentioned as occurring on the opposite side, but always received a negative answer. The W. side was generally from one third to one half as wide as the E. side or right-hand side."

In the tornado investigated by Dr. Anderson, while it had a rapid progressive motion, the left-hand side of the destructive part seems to have been entirely wanting, so far as the prostration of trees, at least, is concerned.

RAINFALL IN TORNADOES.

262. The rapid ascent of nearly saturated air in a tornado must give rise generally to a great amount of rainfall. Some idea of this may be formed from considering the amount of aqueous vapor in the ascending air, drawn mostly from the strata near the earth's surface, and its capacity for vapor, from which the amount left in the air, after it has ascended up to altitudes where it has become cooled down to a lower temperature, becomes known. Suppose the air temperature near the earth's surface is 25° C, but that the depression of the dew-

point is 5°. The tension of aqueous vapor in the air, in millimeters of mercury, is that of saturation at 20°; and with this as an argument Table 11 gives 17.36 mm. The density of vapor being 0.622, this gives $17.36 \times 0.622 = 10.80$ mm. for the height of a column of mercury equal to the weight of aqueous vapor in a homogeneous atmosphere of saturated air at the temperature of 20°. Multiplying this by 13.6, the density of mercury, we get $10.80 \times 13.6 = 146.7$ mm. for the height of a corresponding column of water. Dividing this by 7992, the height in meters of a homogeneous atmosphere, we get 0.0184 mm. for the depth of rain in each stratum of saturated atmosphere of one meter in depth, and of the temperature of 20° if the vapor were all precipitated. If now we suppose that the ascending velocity of the air at the base of the cloud where condensation first commences to be 60 meters per second, then the depth of rainfall, supposing the current to extend so high that all the vapor is condensed, and that the rain all falls directly back, would be $0.0184 \times 60 = 1.1$ mm. per second, or 0.066 m. (about 2.6 inches) per minute. With higher temperatures and higher degrees of saturation the rate of rainfall would be considerably more.

The preceding calculation, however, upon the assumptions made, gives us only a rough approximate idea of the possible rate of rainfall, for the air, at least all of it, never ascends so high that all the vapor is condensed; but, on the contrary, as has been explained in § 241, the ascending current may be deflected off laterally on all sides at no very great altitude. The vapor, therefore, which is carried up mostly by the more rapid currents in the interior of the tornado after being condensed into rain, is carried outward above by the outflowing currents there, and falls mostly over the regions adjacent to the violent part of the tornado, and not directly back over the place where it ascended as vapor. Besides, we have just seen that where there is an ascending current of only moderate velocity, small rain-drops cannot fall back, but are carried up to where the current is outward, and then out where there is little or no ascending current, and where they are permitted to fall to the

earth. In fact, with an ascending velocity even very much less than that assumed above, no ordinary rain-drops by Table VII could fall directly back to the earth.

WATERSPOUTS.

263. As soon as the ascending and expanding air in a tornado cools down to the dew-point corresponding to the diminished vapor tension as it gradually comes under less pressure, condensation and cloud formation take place. Where the air ascends vertically, we have seen, § 27, this takes place at an altitude which, in meters, is equal very nearly to the depression of the dew-point at the earth's surface in centigrade degrees multiplied by 125 or 125 $(\tau - d)$.

But the rate of cooling does not necessarily depend upon the ascent of vapor, but only on the amount of decrease of pressure and of expansion, whether this takes place in the ascent or horizontal motion of air, or in air without any motion. In all cases, to a given amount of decrease of pressure and corresponding increase of expansion there is a corresponding decrease of temperature.

If in the annexed figure we suppose that the horizontal isobaric surface in the undisturbed air, and represented by a hori-

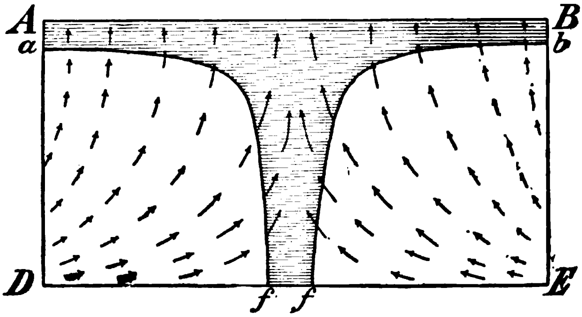


Fig. 3.

zontal line *ab*, is brought down by tornadic action to the earth's surface *DE* at *f* in the form of the line *af*, as explained in § 233, then the air which ascends from the stratum next to

the earth's surface, all supposed to have the same temperature and dew-point, whether it ascends slowly and nearly vertically at the outer border as at *a* and *b*, or more obliquely and rapidly toward and up in the interior, wherever it arrives at the depressed isobaric surface, represented by the curved line *af* or *bf*, it has cooled down to the same temperature; and if the height of that isobaric surface before being depressed in the interior of the tornado was 125 $(\tau - d)$, then condensation and cloud formation take place as soon as the air ascends above or enters within that surface up as high as aqueous vapor is carried by the ascending current. But without and below this surface there is no condensation and cloud formation, and the air remains unclouded. The clouded portion of the air therefore assumes the form of a tapering trunk, as outlined on the two sides by the curved lines *bf* in the figure, and we have the phenomenon called a *waterspout*. A waterspout, therefore, is simply *the cloud brought down to the earth's surface by the rapid gyratory motion of the tornado.* As Espy with a few strokes of the handle of an air-pump produced a cloud in the receiver from the expansion and cooling of the moist air within, so nature, by means of a whirl in the open atmosphere, produces a cloud in the vortex of a tornado, from the expansion and cooling of the air there, on account of the partial vacuum caused by the centrifugal force of the gyrations.

We have seen, § 242, that the air which ascends in the interior of a tornado comes in below mostly near the earth's surface—especially that which ascends where the spout is formed, and so it may be regarded as all having very nearly the same temperature and depression of the dew-point. If the comparatively small amount of air coming in at the sides from the higher strata of the atmosphere had the same depression of the dew-point, it would have to enter some distance within the isobaric surface *af*, *bf* before condensation of the vapor would commence, since having a less pressure before commencing to ascend than at the earth's surface, it would have to arrive when the pressure is less than that of the isobaric surface of incipient condensation for air ascending from the earth's surface, and so

the outline of the spout is determined by the latter. If, however, the air entering from the higher strata should have a much less depression of the dew-point, its vapor might begin to be condensed before entering within and above this surface, and so affect the distinctness of outline of the spout; but this can scarcely, if ever, occur, since the further from the earth's surface, generally, the drier the undisturbed quiet air.

The arrows in the figure represent approximately the air currents coming in from the sides, and mostly near the earth's surface, to supply the draught of the rapidly-ascending current in the interior. It must be understood, however, that these represent merely the currents of the vertical circulation, and that in a tornado there is, in connection with this, also a rapid gyratory motion around the centre. The same system of circulation somewhat takes place in all tornadoes, whether the tornadic violence is sufficient to bring down the spout or not; for, as will be explained, a certain degree of violence is necessary to bring the spout down to the earth's surface. At a considerable distance from the centre, beyond the limits represented by the figure, the air on all sides descends, mostly very gently, and at the same time is drawn in toward the interior.

264. The height of the spout depends upon the hygrometric state of the air, being 125 meters for each degree of the depression of the dew-point at the earth's surface. The diameter of the spout at any given distance l below the level of the undisturbed isobaric surface AB, as at a or f, Fig. 1, § 233, depends upon the amount of gyratory velocity or value of $c = r'v'$, as may be seen in § 234; for the greater the value of c, the greater is the value of r the distance of the isobaric surface from the centre, and, in the case of a waterspout, the greater is its radius. Since the spout is tapering down to the earth's surface, the drier the air and the taller the spout the smaller its diameter at the earth's surface for the same gyratory velocities. For instance, with a small dew-point depression and rapid gyratory motion we get a spout of the form of Fig. 4 in which the cloud is low on the outer border of the tornado where the isobaric surface, or surface of incipient condensation, is not de-

pressed sensibly, and on account of the smallness of the depression l and the largeness of the value of $r'v'$, the value of r is large, as is seen from the inspection of the equation of § 234. With the conditions there assumed of $r'v' = 3000$, and $l = 500$ meters, which corresponds to a depression of the dew-point of 4°, we get for the radius of the spout at the earth's surface,

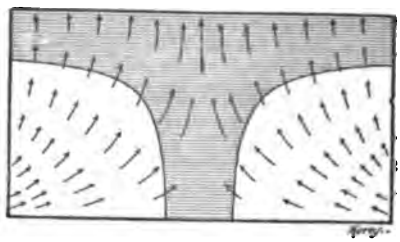


Fig. 4.

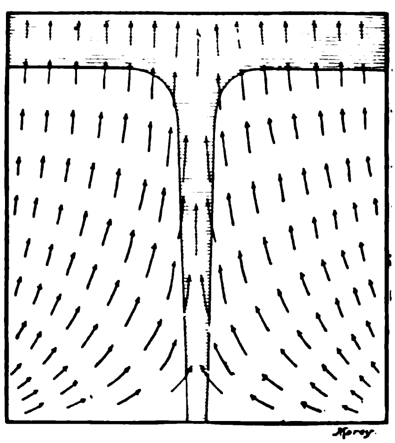


Fig. 5.

$r = 30$ meters, and hence a short spout only 500 meters in height with a diameter of 60 meters at the base. In a tornado under such conditions even the cloud-rim is low, and the dense cloud is brought down to the earth's surface over a considerable area at any given instant, and everything is involved, for the time, in great darkness.

On the other hand, if the air is very dry and the depression,

consequently, of the dew-point very great, we get a very tall spout with small diameter, as represented in Fig. 5. If the depression of the dew-point is 16° C, then the height of the spout is $125 \times 16 = 2000$ meters; and if in this case we put $r'v' = 1000$, we get from the expression of l, § 234, by putting $l = 2000$ meters, as above assumed, $r = 5$ meters nearly, or a diameter of about 10 meters for the spout at the earth's surface. But the value of $r'v'$ cannot be taken so small, or that of l so great, that r vanishes, so that in any case there must be at least a small threadlike spout coming down to the ground according to the theory, in which we take no account of friction.

265. In the preceding results from theory and formulæ in which no account is taken of friction, of course considerable

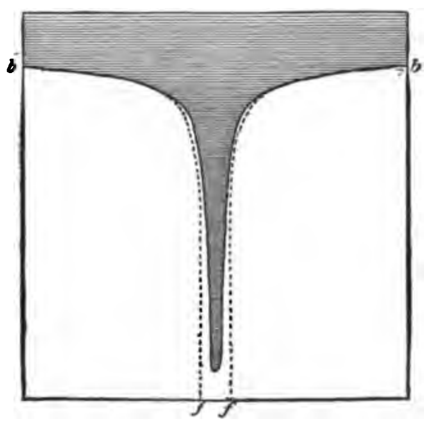


Fig. 6.

allowance must be made for its effect, as has heretofore been pointed out in similar cases. It has been shown that the greater the initial gyratory velocity or value of $r'v' = c$, the greater the diameter of the spout, all other circumstances being the same. Consequently the effect of friction, which tends to diminish the gyratory velocity, is to make the diameter of the spout smaller, and even to vanish where the gyratory velocity is small; and this is especially the case very near the earth's surface where the amount of friction and decrease of gyratory

velocity are comparatively great. For instance, if we had conditions which would give a theoretical spout of the form outlined by the dotted curved lines *bf*, Fig. 6, the effect of friction would diminish its diameter somewhat as represented in the figure, and even prevent its coming entirely down to the earth's surface.

With a still less gyratory velocity, other conditions remaining the same, there would simply be a funnel-shaped cloud as represented in Fig. 7, the gyrations in this case retarded by friction, being merely sufficient to bring the cloud in the vortex a little below the general level of the undisturbed base of the

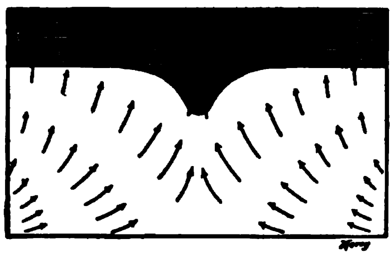


Fig. 7.

cloud. It is in this form it first appears in a tornado, when it is seen at all; and as the energy of the tornado and the velocity of the gyrations increase, it may be brought either wholly or only part of the way down to the earth's surface. Finally, as the energy of the tornado becomes exhausted, and the velocity of the gyrations diminishes, it is drawn up again apparently into the cloud, its last appearance being again that of the funnel-shaped cloud.

In tornadoes of large base with not fully developed and rapid gyratory velocities in the vortex, instead of a completely developed spout or a funnel-shaped cloud, a large basket-shaped cloud is observed suspended from the level of the general base of the clouds. The rapidity of the gyratory velocities in the vortex is not sufficient to bring the cloud down with a sharp point as in the funnel-shaped cloud.

The action of a tornado is somewhat intermittent, now

stronger and again weaker. This is observed frequently in the greater manifestation of violence at some times than at others.

With regard to the tornado force in the S. C. tornadoes of April, 1883, Dr. Anderson states:[68]

"The tracks examined by me did not present continuous lines of destruction, but areas of destruction separated by intervals entirely or almost entirely exempt from destructive forces; from which it is inferred that, while the storm, in its common, ordinary features, pursued its way steadily onward by bodily transference, the tornadic action was developed interruptedly, and progressed by successive transplantings."

That is to say, in the general progress of the tornado, the unstable state and the vertical circulation continues nearly the same, but the gyratory motion depends very much upon the initial and slight whirl which the air near the earth's surface at some distance from the centre may have, as it is drawn into and up the vortex; for while the whirling column above consists somewhat of the same air drifting along, that near the surface, not partaking of the same amount of progressive velocity, is being continually changed, and consequently the value of $c = r'v'$, upon which the gyratory motion depends (§ 232.) According as the value of c is greater or less, for the new portions of air drawn into the central part, is the gyratory velocity and violence of the tornado greater or less. This intermittent action in tornadoes is often indicated by the dropping down and rising up of the spout where the gyratory violence of the tornado is barely sufficient to develop a spout, and is an explanation of this phenomenon; for in that case a very little increase or decrease in the rapidity of the gyrations brings the spout in part or wholly to the earth's surface from the funnel shape, or lets it up again.

Finley states, with regard to the Lee's Summit tornado, that from a certain point "the funnel raised again, committing no damage for a distance of 9¾ miles over an alternation of hill and dale, when it struck a large elm-tree situated in a broad valley with gently sloping sides." While passing over the uneven surface of hill and dale, the gyratory violence, on account of this unevenness, was not sufficient to keep the spout down to the earth's surface, and so here it rose up.

In the account of the Walterborough tornado,[a] it is stated that "its path was continuous for 24 miles, then a gap of 25 miles where no damage was done, followed by a track ¾ mile in length."

With increase of tornadic violence, the spout is brought down to the earth, and the consequent increased destructiveness is attributed to the spout, whereas the latter is simply an effect of the increase of the tornadic violence, and with a drier air there might be no spout accompanying the same amount of violence and destructiveness.

266. From what precedes, it is evident that waterspouts may be of a great variety of forms, varying from that of a cloud brought down over a large area of the earth's surface in a tornado, where the air is nearly saturated with vapor and the general base of the clouds very low, somewhat as represented in Fig. 4, to that which occurs when the air is very dry, and when the tornadic action is barely able to bring the cloud down from a great height into a slender spout of small diameter, somewhat as represented in Fig. 5. Horner says that their diameters range from 2 to 200 feet and their heights from 30 to 1500 feet. Dr. Reye states that their diameters on land, at base, are sometimes more than 1000 feet. Oersted puts the usual height of watesprouts from 1500 feet to 2000 feet, but states that in some rare cases they cannot be much less than 5000 or 6000 feet. On the 14th of August, 1847, Professor Loomis observed a waterspout on Lake Erie, the height of which, by a rough estimate, was a half a mile, and the diameter about 10 rods at the base and 20 rods above."[a]

Judge Williams, in speaking of the tornado of Lee's Summit, where he saw it, says:"[a] "It seemed to be about the size of a man's body where it touched the clouds above, and then tapered down to the size of a mere rod."

The preceding symmetrical waterspouts, deduced from theory upon certain assumed regular conditions, differ of course very much in form from those usually observed in nature, where these regular conditions are rarely found, even approximately. It is assumed in the theory that the progressive mo-

tion of the strata of the atmosphere at all heights to which the tornado extends is the same, from which results a perfectly straight and vertical spout. But this condition usually does not exist, but the upper strata either move faster or slower than the lower ones, and so in the former case the upper part of the spout inclines forward and in the latter backward. Sometimes the progressive motions of the atmosphere in the different strata are not only very irregular at different altitudes, on account of the great abnormal disturbances to which the winds are subject, but likewise very changeable at different times, so that in this case the spout may not only be bent and sinuous, but subject to continual contortions and twistings. Hence, the Rev. S. R. Reese describes the appearance of the spout of the Lee's Summit tornado, as he saw it," to be "that of a funnel, at times elongated in a serpentine-like form of heavy mould, hung up by the head and writhing in agony, its tail curling and lashing as if actuated by the impulses of a living body. It was above the ground and doing no damage, though at times, in its violent contortions and struggles, it would descend very near the ground."

Other irregularities arise, also, from unequal distributions of temperature and of aqueous vapor in all portions of the air around on all sides. The base of the cloud also, from which the spout depends, does not have the smoothness and the uniformity on all sides shown in the preceding graphic representations of the results deduced from theory. For although the surface of incipient condensation, as the air, charged with vapor, ascends, may be somewhat regular, yet the cloud above is at some distance from the vortex brought down below this level in the form of scud clouds, which often gives it an irregular and jagged appearance. In fact, this regular base of incipient condensation and the extreme upper wide part of the spout or funnel is frequently concealed by such clouds, which are brought down and, before they are evaporated, are carried into the vortex. For although the air is drawn into the vortex below, at the earth's surface mostly, as has been explained, yet it also seems to be drawn down also, at no great distance

from the vortex, to supply the ascending current; and so it brings portions of the cloud above down, sometimes even near to the earth's surface. Thus, in the Lee Summit tornado,"

"The funnel is represented as reaching the ground, which under such circumstances causes the narrowest part to be some distance above the surface, but when it rises the lower portion tapers down to a very small diameter. Dark masses of cloud shot downward on either side of the funnel, entering it just above the ground, apparently thereafter rushing upward through the centre."

Espy says:

"Whenever low clouds appear under the rim of the hurricane cloud, they always are seen in the form of scud moving rapidly toward the centre of the great hurricane cloud, where they unite with its base and ascend with the ascending current, where the barometer is very low."

267. Waterspouts, especially on land, frequently have the appearance of a widening at the bottom on the earth's surface. On land, dust and a great many light substances are carried up in the interior; and as they are being collected from all sides on the earth's surface by the inflowing currents, which are here more nearly radial, toward the vortex below, they often assume the form of a cone, which, in the first formation of the spout, seems to rise up and meet the descending spout falling apparently from the clouds, and thus the whole phenomenon often assumes the form of an hour-glass. Of the great tornado of West Cambridge (now Arlington), August 22, 1851, it was said:

"To some who watched it closely its form resembled a tall, wide-spreading elm tree. To others it appeared like an inverted cone. Several represented it as a dense upright column, and a few as having the shape of an hour-glass."

The several observers no doubt saw it at different times and under somewhat different circumstances in its progressive motion over the inequalities of surface. It was only where the earth contained a great amount of dust, or other light materials on the surface, that the right cone at the base was observed, and where the whole assumed the form of an hour-glass.

With regard to the Lee's Summit tornado, it was stated:[11]

"The small funnel cloud was seen reaching part way to the ground; at the same time an inverted funnel of dust and light materials formed over the earth beneath and reached up to it."

The graphic representation of it in Fig. 8 is given in the account of it.

Sometimes the right and inverted cones do not come together, but a small space is left between the two vertices of the

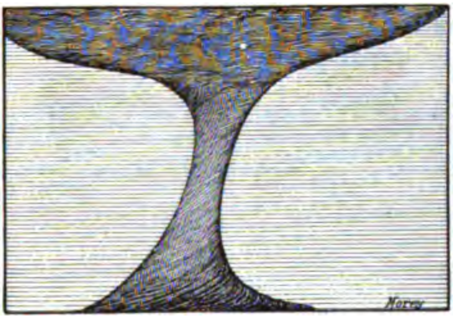


Fig. 8.

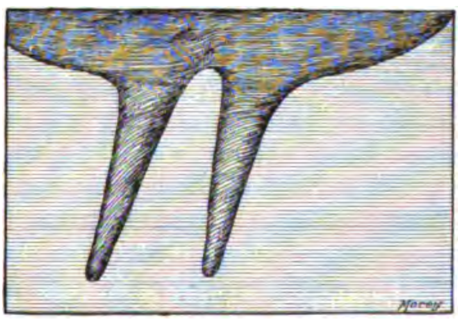


Fig. 9.

cones, pointing the one down and the other up toward each other. From no accounts can it be inferred that in land tornadoes the lower part is composed of condensed vapor—at least the outer part of it, which causes the widening, but the true spout may reach through it to the earth's surface.

There are often two or more spouts in close proximity. The graphic representation in Fig. 9 of two spouts, brought part of the way down to the earth, seen by Mrs. Kerr in the Lee's Summit tornado, is given by Finley. Sometimes several small spouts protrude from the lower base of the cloud in the same vicinity, some to a greater and some to a less distance down, sometimes not differing much in size; at others there is a larger and principal spout accompanied by one or more smaller ones. The accompanying representation of a group of little spouts or funnels, as seen by McLaren at one time in the Delphos tornado, is given by Finley.

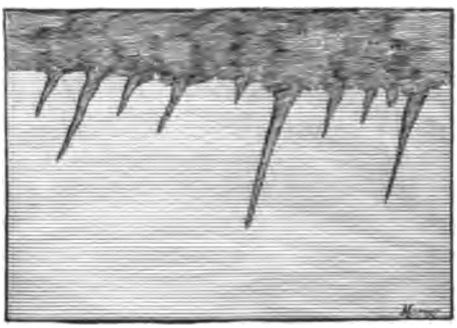


Fig. 10,

As the tornado originates in air in the unstable state, it often happens that there is about an equal tendency in the air of the lower stratum to burst up through those above at several places in the same vicinity at the same time. Each of these gives rise to a separate and independent gyration in the atmosphere, and a small funnel where they are of sufficient violence; but generally, as they increase in dimensions and violence, they interfere with one another and finally become united into one. This seems to have been the case in the tornado of St. Cloud and Sauk Rapids, in the formation of which it is stated that as the clouds began to revolve sharp points suddenly shot downward. As in cyclones there are generally smaller secondary ones included, complicating the resultant motions of the air and causing irregularities in the isobars, so in tornadoes

there are undoubtedly secondary whirls independent of the main one, though not always accompanied by corresponding funnel clouds visible above. This seems to be indicated by side or spur tracks sometimes observed in connection with the main track, and by isolated spots at some distance from the main track "with every tree uprooted and piled in confusion."[66]

268. Waterspouts at sea are usually more regular and better defined than those on land, and the whole area of tornadic disturbance is generally smaller, so that the spouts may be approached with safety within a very short distance, and it is only the larger and more violent ones that seriously injure a ship running into them. The destructive gyratory winds, even in the larger ones, extend only a short distance from the centre, and at distances a little greater scarcely a breeze sometimes is experienced. The reason of this is that, the surface of the sea being smoother than that of the land, there is a more nearly perfect development of the gyrations, and a greater concentration of energy in the centre of the vortex, although the whole amount of energy is generally smaller on sea than on land, since this arises from the unstable state, which is more liable to occur, and to a greater degree of unstability, on land, where the surface of the earth becomes much warmer than that of the ocean. The whole disturbance, however, is simply a tornado—it may be a very small one, with the phenomenon of the waterspout developed in the same way as in a tornado on land.

It was formerly supposed that the spout consisted of water drawn up into the clouds from the sea, and that the real water-spout was found on seas and lakes only, and hence the name. It is true that a considerable amount of water may be drawn up from the sea, but this is merely an incidental and secondary matter and has nothing to do with the formation of the spout. The amount of water drawn up is so small generally in comparison with the amount of rainfall, that the latter is never observed to be sensibly affected by it at sea, but always appears to consist of fresh water.

When we consider how houses, men, and various kinds of

heavy bodies and *débris* are drawn up into the vortex of land tornadoes and thrown out above in all directions, as explained in § 258, it is not surprising, but to be expected, that much water and other things would be carried up in tornadoes at sea. Here the very centre of the vortex may become nearly a vacuum, and hence the tendency of the water from the surrounding great pressure would be to rise up nearly 32 feet; and having risen so far, or at least to a considerable height, the rapidly ascending currents would carry it farther, and, being lashed and separated into drops, it would then be carried to still great heights, as raindrops are in the interior of a tornado.

We have accounts of not only water, but of fish and frogs being drawn up in the vortex of small tornadoes from ponds and creeks, leaving, for the moment, the bottoms of the latter dry, and afterwards scattering the fish and frogs over the neighboring land. Tomlinson says:[30]

"Showers of fish and frogs are by no means uncommon, especially in India. One of these showers, which fell about 20 miles south of Calcutta, is thus noticed by an observer: 'About two o'clock P.M. of the 20th inst. (Sept. 1839) we had a very smart shower of rain, and with it descended a quantity of live fish, about three inches in length, and all of one kind only. They fell in a straight line on the road from my house to the tank, which is about 40 or 50 yards distant. Those which fell on the hard ground were, as a matter of course, killed from the fall; but those which fell where there was grass sustained no injury, and I picked up a large quantity of them, 'alive and kicking,' and let them go into my tank. The most strange thing which ever struck me in connection with this event, was, that the fish did not fall helter-skelter, everywhere, or here and there, but they fell in a straight line, not more than a cubit in breadth."

The following is taken from an account of a waterspout given by Sidney B. J. Skirtchly, H. M. Geo. Survey:[76]

"During the summer of 1870, while in Deeping Fen, on a day when the wind was blowing in gusts, carrying the dry powdery peat-dust in clouds before it, I observed a whirling column of dust advancing toward me. It was like those small pillars so frequently seen in streets of a town on such a day, but it was considerably larger, being from 15 to 20 feet in height. When it was first seen it was advancing from the far side of a *ground*, as the uninclosed fields are called, toward me at the rate of

about six miles an hour, and was distant some 500 yards. It moved with an unsteady, staggering motion, accompanied with a rushing noise. I stayed to watch it across a dike about 15 feet wide, which ran directly across its path. The smaller dikes it seemed to cross without affecting them; but on reaching the one in question, it whisked the water up into a waterspout some ten feet high with a gurgling, hissing sound, and steering directly across the dike burst, on reaching the opposite shore, projecting a considerable quantity of water upon the land. This effort seemed to spend its force, for the dust-column resumed on the opposite land was but small in proportion, and after swaying about for a few yards, died away."

This was simply a whirlwind, as small tornadoes are generally called, in which the air was too dry and the energy too small to develop the real waterspout of particles of condensed vapor, yet it seems to have had considerable power in raising up water from the dike.

In addition to the water carried up in the spout of a tornado there is often also much spray about the base of the spout which sometimes assumes the form somewhat of the sand and dust cone in a land tornado, causing an apparent widening of the spout at the base.

269. Small waterspouts observed on seas and lakes in clear, calm, and hot weather usually arise from a state of unstable equilibrium in the lower strata of the atmosphere. In such cases the whirling of the air and the agitation of the water is first observed below, and afterwards the formation of a cloud above, and finally the complete spout, unless the atmosphere is very dry. When, however, the air is very moist,—near the point of saturation,—such spouts extend up to only a small height, and are not accompanied by any rainfall, since they are usually too small and continue too short a time to send up the vapor to so great a height that it can be condensed and collected into drops and fall as rain.

A number of such small spouts are often seen at one time in the same vicinity. When the air is brought to the unstable state it is liable to burst up through the strata above at several places at nearly the same time, and then, if there is any whirling motion, even only very small and imperceptible, as in the

case of the water in a basin, there is a concentration of this motion into rapid gyratory motions at the centre of each one of these upbursts, which may continue to increase until a waterspout is formed.

M. Defranc" has given an account of a great many waterspouts of the class which occurs mostly in weather which is nearly clear and calm. He says:

"On the 23d of June, 1764, a waterspout was seen on the Seine, which had its base on the river and reached up into the clouds. It was judged to be about three feet in diameter where it touched the river. There were some parts transparent, which allowed the ascension of the water to be seen. It finally broke at about one third of its height. The lower part fell in rain, the upper part was drawn up into the cloud in a second of time and the phenomenon was followed by hail."

This seems to have been a tall spout notwithstanding the smallness of the diameter. The air in the vicinity must have been nearly calm, and in the unstable state up to a considerable altitude. Being a very tall, slender column with its base upon the river, the friction was small, and so it approximated to the theoretical case of no friction in which the least amount of initial gyratory motion would bring down to the earth a fine thread-like column of very much rarefied and cooled air, which gives rise to a slender waterspout. At the time when the air in the central column is not quite rarefied and cooled sufficiently to condense the vapor, the least increase in gyratory velocity brings it into this condition, and then the spout darts down from the cloud above almost at once. Just the reverse of this takes place in the case of a slender spout when the gyrations are a little decreased from any cause, and this spout seems to have been thus drawn up. The lower part which is said to have fallen as rain was, no doubt, simply the water raised from the river, as fish and frogs are said to be, and not rain. The falling of hail indicates that the ascending currents extended very high up into the upper and very cold strata of the atmosphere

Defranc also states:

"On May 17, 1763, Captain Cook saw six waterspouts on Queen Charlotte Sound. In one of them a bird was seen, and in arising was drawn in

by force and turned around like a spit. Their first appearance was indicated by a violent agitation and elevation of the water. When the tube was first formed or became visible, its apparent diameter increased. It then diminished and became invisible at its lower extremity."

The violent agitation and heaping up of the water before the spouts appeared show that the gyrations and barometric depressions in the centre existed before the spouts became visible, and that the spouts appeared only after the diminution of tension and of temperature became sufficient to condense the vapor. The fact of the bird's being drawn in and whirled around shows that the air is really drawn in from all sides and up in a spiral in accordance with theory.

The frontispiece to the title-page is the view of a waterspout as observed off the coast of Sicily, and sketched by Mr. Morey in August, 1876. The following is his account of it:

"It formed during a comparatively calm and clear afternoon, with the temperature of the atmosphere not much, if any, higher than usual for that latitude and place. The wind was so light as to barely cause a slight ripple on the surface of the sea, but sufficient to give the spout a perceptible motion from east to west. Three little protuberances first appeared under the cloud, the centre one rapidly elongating connected with the water. The two on the sides increased and decreased in length, but never reached further down than one fifth the length of the main spout.

"It lasted about twenty minutes, and disappeared shortly after the breaking of the main column at a point which seemed the connection between the funnel and the water.

"No noise accompanied the break in addition to the noise resembling a cataract, which lasted from the time the spout connected with the water to the break up. No electrical phenomena were noticed at all.

"A rough estimate of the main column gave its height between 700 and 800 feet.

The stability of the formation during the twenty minutes of its duration was wonderful in the extreme."

Of the two partial spouts or funnel-shaped protuberances on each side the one, according to the sketch, seems to have caused a considerable agitation of the water beneath it.

270. Waterspouts of the class here considered are very often seen along the eastern coast of the United States in the vicin-

ity of the Gulf Stream, not only during the summer season, but likewise in the winter. On the Little Bahama Bank as many as fifteen have been seen at the same time. The following account of them is given by an officer of H. M. Surveying Vessel "Sparrow-Hawk," employed in the West Indies:"

"I have noticed that the first movement which eventually produces a waterspout is a whirlwind on the surface of the water, gradually increasing in velocity of rotation and decreasing in diameter as it travels along before the prevailing wind. The spray is lifted up to a height of from five to ten feet, and then gradually melts away, assuming the appearance of hot air, which is visible (still rotating) to a similar height above the spray. A motion amongst the clouds soon becomes apparent, a tongue is protruded, and the spout becomes visible from the top downwards.

"On one occasion a portion of a spout appeared for a moment in mid-air above the disturbances on the surface of the water.

"Although these appearances are commonly called waterspouts, I have been informed by men who have been caught in them that they contain no water and should be properly called 'windspouts.' The small fore-and-aft-rigged schooners that ply on the bank do not fear them, although a prudent captain would probably shorten sail to one. I have been unable to hear of an accident having occurred through a vessel being caught in a waterspout.

They frequently cross the land, but no water falls; they take up any light articles, such as clothes spread out to dry, straw, etc., that happen in their course, but have never been known to carry anything with them to a distance."

It is seen from the preceding account of this class of spouts that the whirlwind, and the consequent agitation of the water beneath, first commence and gradually increase until the strength of the gyratory motion becomes sufficient to bring down the spout, and that this first appears above as a funnel or short spout, and then becomes visible from the top downward, the same as in the larger and more violent land tornadoes. It seems that there is no water carried up in these small whirlwinds except some spray to a short distance; but in the larger ones, and especially the land tornadoes, some water is evidently carried up when they pass over water. If it were necessary to change the name, which, as in many other things, was given before the thing was understood, it would be

more appropriate to call them *vapor-spouts*, since they are evidently composed of condensed vapor, and no amount of rotary motion in dry air would produce such a phenomenon.

Where the atmosphere is clear and calm over a river or small lake, as may frequently happen where it is surrounded by highlands and forests, and the air is very nearly saturated with vapor, very diminutive spouts of small altitude are sometimes seen. Under such conditions the air near the surface of the water becomes heated, both by the direct and reflected rays of the sun, and a stratum of air of small depth next to the surface is brought to the unstable state, in which little whirlwinds and ascending currents originate in the usual way. The air, being nearly saturated, would have to ascend to only a small altitude without any whirl before cloud formation would take place, and this being so near the earth's surface, a slight whirling of the air is able to bring the cloud down to the surface of the water in the form of a little spout.

271. Such small spouts have been observed, under very peculiar circumstances, to be hollow. M. Boué,[79] in the year 1850, observed three small waterspouts at the same time on Lake Janina, from the top of a high mountain. The weather was entirely clear, without clouds or wind, but very oppressive and hot. The spouts seemed to rise up from the lake, and he could look down into the top of them and see that they were hollow in the middle. The same seems to exist in some measure in large spouts both on sea and land, as is indicated by the central part of the column appearing lighter than the surrounding parts.

This hollowness, or less density of the condensed vapor, in the middle, arises, no doubt, from the effect of the centrifugal force of the very rapid gyratory velocity near the centre, in driving the condensed vapor, as soon as formed, out from the centre, and also from the fact that the central part is so rarefied and cooled down that but little vapor, even before condensation, can approach near it, since it is mostly condensed before arriving there; just as in the vertical ascent up into the region of freezing cold very little uncondensed vapor is left after

having ascended to that altitude. The central part of a water-spout corresponds in temperature, density, and capacity for vapor with these high dry strata of the atmosphere in the surrounding vicinity, as is understood from Fig. 1, § 233.

In fact there seems to be a slightly analogous phenomenon even in cyclones, in which the clouds are frequently less dense, and even entirely disappear sometimes, in the central part, giving rise to what is called "the eye of the storm," though the explanation is only in part the same.

HAIL-STORMS.

272. A hail-storm is simply a tornado in which the ascending currents are so strong, and reach so high up into the upper strata of the atmosphere, that the rain-drops are carried up into the cold regions above, and into the central part within the isobaric and isothermic surface of the freezing-point, where they are frozen into hail. In fact hail, as well as rain, is almost a universal accompaniment of a tornado, and so, as the latter, it is usually found in the S.E. quadrant of a cyclone and at a considerable distance from the centre, as has been shown to be the case in Russia by Klosovsky from observation (§ 311). Finley says,[3] in his account of the tornadoes of May 29 and 30, 1879:

> "In referring to the several storm descriptions, it will be found that rain and hail invariably preceded the tornado cloud from ten to thirty minutes, nearly always attended by a southerly wind."

But this applies mostly to the more violent tornadoes, such as were investigated by him, and not to every class of atmospheric phenomena which we have included under the generic name tornado.

It has been shown in § 233 that every isobaric and isothermic surface, however high above the earth's surface, would in a perfectly regular tornado, and in the case of no friction, be brought down by the centrifugal force of the gyratory motion to the earth's surface near the centre of the tornado, as represented in Fig. 1. The horizontal and isothermic surface,

therefore, high up in the air, which separates the strata of freezing temperature above from those of a non-freezing temperature below, would, in case of no friction, be brought down by tornadic action to the earth's surface at *c*. The effect of friction, however, prevents its bringing the freezing atmosphere entirely down, or rather its so expanding and rarefying the air

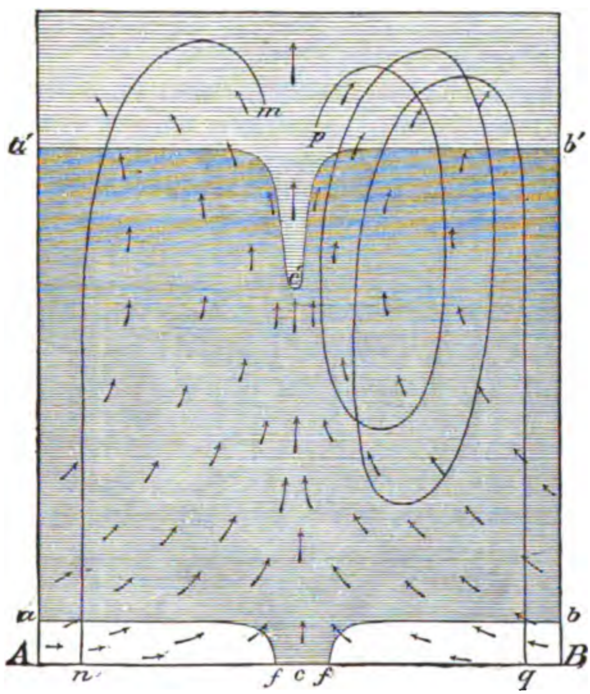


Fig. 11.

in the central part that the air in consequence is reduced to the freezing temperature down to the earth's surface. But still the air of freezing temperature is brought down in the central part in the form of a funnel, considerably below the general level of its base when undisturbed, just as it has been shown the cloud region is, Fig. 7, and as is represented by $a'c'b'$, Fig. 11, in which the unshaded part represents the unclouded air below and beyond the surface of incipient condensation, here

supposed to be brought entirely down to the earth's surface *A B*; the darker shaded portion, the part in which the vapor is condensed to ordinary cloud; and the lighter shading above, the part where the vapor is converted directly to snow, and the rain-drops, carried up by the ascending currents, to hail. The height of the surface of incipient freezing, or temperature of 0° C., where the temperature of the air at the earth's surface is given, may be ascertained in the same way as the temperature at any given height, as explained in § 27.

273. In a torrado the air at and near the earth's surface is drawn in and ascends on all sides until it arrives at the surface of incipient condensation, represented by the lower line *ab*, brought down to the earth on each side of the centre at *f*, at which surface, as has already been explained, cloud-formation commences. In its further ascent and approach toward the interior central part the vapor contained in it is gradually condensed, and the temperature of the air reduced, until it arrives at the surface of incipient freezing, either up at the undisturbed level beyond the influence of tornadic action as at *a′* or *b′*, or where that surface is brought down below this level near the centre in the region of *c′*. As the air, with the uncondensed vapor yet remaining, enters above and within this surface, the vapor left is gradually frozen to snow, and the higher it ascends the more of it is so changed. This snow, however, never falls to the earth in the summer season, but melts soon after falling below the level of incipient freezing. In the winter season, at times when this level is not very far above the earth's surface, it reaches the surface as soft snow, where there is no tornadic action and only gentle ascending currents.

In the ascending current of a tornado, as in that of the equatorial calm-belt or of a cyclone, the rain-drops are formed down in the cloud region and carried upward until they become too large to be supported by the current, and so fall to the earth, as explained in § 111. In a tornado, however, the ascending current is often so strong that the rain is supported until, by the blending of the small drops by coming in contact,

very large drops are formed, and the strong ascending currents often extend so high that these large drops are carried away up into the region of freezing temperature, represented by the upper lighter shading in the preceding figure. They are there frozen, and after having been carried up and outward above to a distance from the centre where the ascending current is not strong enough, by the formula of § 247, to keep them up, they slowly descend, and, receiving additions of ice as they fall, as long as their temperature remains below zero, from the rain below, which is being either kept suspended in the air or being carried farther up, they finally fall to the earth as *solid hail-stones.*

274. But the origin of the hailstone is often not a rain-drop, but a bunch of snow formed in the snow region, and moistened by the rain carried up into this region before it has had time to become frozen into hail. This moist snow is kept up there until it freezes, and after that, while being kept up there, and after it commences to fall, as long as its temperature remains below zero it continues to receive a coating of ice from the rain which is carried up past it and that through which it falls. As it grows and is carried outward above where the ascending current is weaker, it finally becomes too heavy to be kept suspended in the air, and it falls to the earth a hailstone, with a *kernel of frozen snow in its centre.*

In such a case we may imagine the soft ball of snow to have originated in the snow region at *m* and then to have been kept partially suspended and to have been carried up and out slowly from the centre to a distance where it could drop down, having become meanwhile reduced to a low temperature, and that on its way down it received a coating of ice from the small ascending rain-drops, or others with which it came in contact during its fall, and that it finally reached the earth at *n*.

It often happens, however, that in falling very gently where the ascending current at no great distance from the vortex is not quite but nearly sufficient to keep it suspended, it is drawn in again by the indrawing currents from all sides toward the vortex, as the scud clouds under the rim of the tor-

nado cloud, § 266, where the ascending current is sufficient to carry it up again into the snow region, where it receives a coating of snow moistened by the small rain-drops carried up into the snow-region before they freeze. This coating now becomes frozen solid, and the whole mass, it may be, is reduced considerably below zero temperature before it is carried out above, where it can gradually drop down again toward the earth; and in falling, even through the lower part of the snow-region where there is little snow, but mostly rain-drops not yet frozen, it receives another coating of solid ice; for its temperature having been reduced considerably below zero, it continues to freeze the rain coming in contact with it for a long distance after having passed down into the cloud region. But in gently falling down it may be drawn in a second time toward the centre and be carried up by the ascending current into the snow-region and receive another coating of wet snow over the last one of solid ice, and in falling receive another coating of solid ice in the same manner as before. This process may be repeated a number of times, in each of which the hailstone, disregarding its gyratory motion all the while, describes a kind of orbit, not returning into itself, somewhat as represented in Fig. 11, until finally it is carried out above so far from the centre, or the strength of the tornado becomes so much weakened, that it is no longer carried in toward, and up in, the central part, but falls to the earth a *hailstone with a snow-kernel and a number of alternating concentric coatings of solid ice and frozen wet snow.*

In this case we may suppose the small snowball to have originated at *p* and to have been carried up and out, and then to have fallen down and to have been drawn in toward the centre several times, until finally it was carried out so far, and had grown to be so heavy, that it was not brought in toward the centre but dropped down to the earth at *q*, Fig. 11.

Hail-storms occur mostly in the summer season. In the winter season the plane of incipient freezing is so low that there is little or no rain-cloud region in which large rain-drops can be formed and in which, after having been frozen above,

they are increased in size in falling through. Winter hail, therefore, consists of only very small particles of frozen water.

275. A remarkable example of hailstones with concentric coatings was observed at Northampton, Mass., June 20, 1870, an account of which has been given by Mr. Houry.[85] In this case hailstones fell weighing over a half pound, and most of them were formed of concentric layers, like the coats of an onion. Mr. Houry states that in one of them he counted thirteen layers, indicating, as he says, that it had passed through as many strata of snowy and vaporous clouds. The true explanation is, that it oscillated as many times between the rain-cloud and the snow-cloud regions, or, in other words, that it performed six or seven revolutions with the lower part of its orbit in the rain-cloud, and the upper part in the snow-cloud, in the manner represented in Fig. 11.

In corroboration of the preceding theory of the formation of hail, and as an exemplification of the motions of a hailstone, as represented above, the following extract from a letter of Mr. John Wise, the balloonist, written to the New York *Tribune* in Feb., 1857, with reference to an ascent which he made in a tornado, is given here, from which it is seen that he described similar orbits with his balloon in the tornado cloud:[86]

"This storm originated nearly over the town of Carlisle, Pennsylvania, on the 17th of June, 1843. I entered it just as it was forming. The nucleous cloud was just spreading out as I entered the vortex unsuspectingly. I was hurled into it so quickly that I had no opportunity of viewing the surroundings outside, and must therefore confine this relation to its internal action. On entering it the motions of the air swung the balloon to and fro, as around in a circle, and a dismal, howling noise accompanied the unpleasant and sickening motion, and in a few minutes thereafter was heard the falling of heavy rain below, resembling in sound a cataract. The color of the cloud internally was of a milky hue, somewhat like a dense body of steam in the open air, and the cold was so sharp that my beard became bushy with hoar-frost. As there were no electrical explosions in this storm during my incarceration, it might have been borne comfortably enough but for the seasickness occasioned by the agitated air-storm. Still, I could hear and see, and even smell, everything close by and around. Little pellets of snow (with an icy nucleus when broken) were pattering profusely around me

in promiscuous and confused disorder, and slight blasts of wind seemed occasionally to penetrate this cloud laterally, notwithstanding there was an upmoving column of wind all the while. This upmoving stream would carry the balloon up to a point in the upper cloud, where its force was expended by the outspreading of its vapor, whence the balloon would be thrown outward, fall down some distance, then be drawn into the vortex, again to be carried upward to perform the same revolution, until I had gone through the cold furnace seven or eight times; and all this time the smell of sulphur, or what is now termed ozone, was perceptible, and I was sweating profusely from some cause unknown to me, unless it was from undue excitement. The last time of descent in this cloud brought the balloon through its base, where, instead of pellets of snow, there was encountered a drenching rain, with which I came into a clear field, and the storm passed on."

The "cold furnace" through which he passed seven or eight times was most probably in and near the interior and upper part of the vortex, in the vicinity of *c'*, Fig. 11, where the snow-cloud and cold upper strata are brought down in the form of the funnel-cloud in tornadoes, or, it may be, still lower in the form of a waterspout, for it must be borne in mind that the balloon could be carried into this snow-region, or at least into a region of very low temperature, without ascending to very high altitudes by being carried close into the centre where this region is brought down. Low temperatures are reached by going in toward the centre as well as by ascending to higher altitudes.

The small pellets seem to have originated in small rain-drops which, being carried up into the snow-cloud, received a coating of snow, but had not received a subsequent one of ice when they struck his balloon.

The motions of the balloon in which M. H. Lecocq recently ascended from Paris seem to have been similar, on entering up through the base into a thunder-cloud, to those of Mr. Wise's balloon. According to the account:[1]

"The balloon, influenced by electrical attraction, rose toward the cloud, accompanied, or rather preceded, by the pieces of paper which the balloonists had thrown from their basket. At 20 minutes of 8, and at a height of 1100 meters, the balloon entered a cloud-mass of greenish-gray color, which immediately shut out from them all sight of the

earth. . . . The balloon constantly rotated and ascended and descended without the interference of the balloonists; and, what is a very rare thing in a balloon, they felt almost constantly a very considerable wind, which shook the balloon and gave to the basket a swinging motion of considerable amplitude."

The account further states:

"At certain times a sensation as of a current of cold air was very perceptible. This was followed immediately by a rapid ascension, and the expelled gas descended even to the basket. During one of these ascents the balloon reached a height of 1600 meters, which was the maximum."

The balloon was not drawn into the cloud by electrical attraction, but carried up into it by the ascending currents, which gave rise to the cloud, and with a velocity less than in the case of the pieces of paper, because the latter had more surface exposed to the ascending currents, in proportion to their mass, than the balloon. At the height of 1100 meters the vapor began to be condensed to form the base of the cloud. The relative air-current in this case arose from the air's ascending and being drawn in obliquely toward the centre past the balloon, which was acted upon by the force of gravity and the centrifugal force of the gyrations. The balloon ascended rapidly when drawn in near the centre where the ascending current is rapid, and descended when carried out about [illegible] where the ascending currents are weak. It was near the [illegible] centre where the sensation of a cold current was experienced, and then the balloon ascended rapidly, because [illegible] ascending current is strong.

In this thunder-cloud there was no doubt [illegible] tornadic action, as indicated by the observed [illegible] and ascent and descent of the balloon, though [illegible] give rise to either hail [illegible] spout.

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increases most on the edges. Others are of a pyramidal form, or of an irregular oval shape with uneven surface. They also appear sometimes to be fragments of large pieces of ice, broken by collision with one another by the gyratory vortex of the tornado.

In Iowa, in 1880, according to Dr. Hinrichs,

"Hailstones 12 inches in circumference have been measured in Sac County. In Davis County a flattened disk of hail measured 4½ inches in diameter, and was 2 inches thick. In Iowa County a group of ice crystals fell, 2 inches in length, 1½ inches wide, and 1 inch thick."

From the report on the tornadoes of May 29 and 30, 1879,[*] have been extracted the following notices of some very extraordinary hail-falls:

In the Lee's Summit tornado, Mrs. Irwin stated that "hail like 'large chunks of ice,' with very little rain, fell just after the storm-cloud passed."

In the Delphos tornado, it is stated:

"On the farm of Mr. Peter Bock, in the adjoining township of Fountain, about 4 miles W. of the storm's centre, and during the hail-storm that preceded the tornado, masses of ice fell as large as a man's head, breaking in pieces as they struck the earth. One measured 13 inches in circumference, another 15, and a hole made by one that fell near the place of Mr. J. H. Kams measured 7 inches across one way and 8 the other. This immense fragment of aerial ice broke into small pieces, so that its exact size could not be determined."

It is also stated in the copy of a letter from Mr. Billingsley, of Delphos, that several hailstones fell which measured 12 to 15 inches in circumference and 1 to 2 inches in diameter. These must have been in the form of disks the thickness of which was from 1 to 2 inches.

In the account of the Lincoln County tornado, it is stated:

"At first the hailstones were about the size of marbles, but they rapidly increased in diameter until they were as large as hens' eggs and very uniform in shape. After the precipitation had continued about fifteen minutes, the wind ceased and the small hail nearly stopped, when there commenced to fall perpendicularly large bodies of frozen snow and ice, some round and smooth and as large as a pint bowl, others inclined

to be flat, with scalloped edges, and others resembled rough sea-shells. One of the latter, after being exposed an hour to the sun, measured 14 inches in circumference."

CLOUD-BURSTS.

277. Considering the strong ascending currents in tornadoes, and their great supporting power, as deduced from theory and exemplified in numerous cases of observation, it is not to be wondered at if great accumulations of rain and hail are sustained for a while by these currents at a considerable altitude in the vortex of the tornado, and then, on account of a sudden weakening or breaking up of the tornado for some reason, a great amount of rain and hail should sometimes fall to the earth in a short time. Such abundant and sudden precipitations are called *cloud-bursts*. If the velocity of the ascending current in the interior is not so great that the rain is all carried up where the current is outward from the vortex, or where this outward current and the centrifugal force of the gyrations together are able to drive it out where it can fall through the weaker ascending current, and yet is great enough to prevent its falling back, then in the whole of the lower part of the cloud in the central part of the tornado, even up to an altitude of three miles or more, there may be a great accumulation of rain, prevented by the ascending current from falling, and also by the centripetal in-drawing current below a given level from being carried out and dispersed. Of course, a considerable part of the energy of the tornado is required to support this mass, so that, as this increases, the strength of the ascending current is weakened, until finally the whole mass suddenly falls. Or the whole system may become weak and break up from some other cause, when the same result follows. This is especially liable to occur in mountainous regions; for if we suppose that a tornado thus heavily charged with rain is moving toward the side of a mountain, its coming in contact with it would interfere very much with the gyrations and energy of the tornado, and tend to break up the whole system almost at once, and let the whole accumulation of water drop

suddenly down. Hence cloud-bursts most frequently occur on mountain-sides.

278. The water in cloud-bursts does not generally fall as rain, but is *poured down*. Long before the ascending current is so reduced as to allow it to fall in drops, the water seems to collect together at certain places and to force its way downward, through the ascending current, in a stream. This it would naturally do, since we cannot suppose that it is ever evenly distributed over any given place, or that the velocity of the ascending current is exactly the same at all points in the same vicinity. A considerable body of water having been collected at certain points, it is there able to force its way down, and it draws into its train much more from all sides on its way, so that continuous streams of water are formed and kept up for several minutes, during which an immense amount of water falls on a number of small spots, while not even rain-drops fall between on account of the strength of the ascending current, through which water can only be poured down. Having collected in large masses and once made an opening for itself at one or more places, the velocity of the streams is gradually accelerated, since the ascending current then is not able to support them, so that on reaching the earth the velocity may become immensely great, and the streams strike with great force. Each one of these descending streams may make a great hole or basin in the ground; and on a steep mountain-side, if the stream continues for a short time only, it may give rise to a mountain slide, or at least to a great ravine, and carry rocks and trees with it down the mountain-side.

279. Immediately after the great tornado at Hollidaysburg, Pa., on the 19th of June, 1838, Espy[34] visited the vicinity and examined the sides of the ridges and mountains. He found a great many holes eight or ten meters in diameter and one or two in depth, according to the nature of the soil and depth of the rock, and the sides often cut almost perpendicularly down on the upper sides, but entirely washed out below, so as to form the commencement of a ravine. Sometimes the current descending through the atmosphere seemed to strike the earth

with so great a force that it made a great hole or basin and then rebounded so as not to strike the earth again on the mountain-side for a considerable distance below. Several of these holes were often found close together in the same vicinity, indicating that the water was poured down at the same time through several openings made in the current of ascending air by the concentration of large bodies of water at these places.

With regard to a ridge a half-mile west of Hollidaysburg, he says:

> "On examining the northern side of this ridge, large masses of gravel and rocks and trees and earth, to the number of 22, were found lying at the base on the plain below, having been washed down from the side of the ridge by running water. The places from which these masses started could easily be seen from the base, being only about 30 yards up the side. On going to the head of these washes, they were found to be nearly round basins from 1 to 6 feet deep, without any drains leading into them from above. The old leaves of last year's growth, and other light materials, were lying undisturbed above, within an inch of the rim of these basins, which were generally cut down nearly perpendicularly on the upper side, and washed out clean on the lower. The greater part of these basins were nearly of the same diameter, about 20 feet, and the trees that stood in their places were all washed out. Those below the basin were generally standing, and showed by the leaves and grass drifted on their upper side how high the water was in running down the side of the ridge; on some it was as high as 3 feet. It probably, however, dashed up on the trees above its general level."

In an account of a remarkable storm which occurred at Catskill, July 26, 1819, it is stated that "the rain at times descended in very large drops, and at times in streams and sheets." In the mountainous regions of the States and Territories of the United States immediately east of the Rocky Mountain range, where tornadoes especially abound, the immense amount of water which sometimes drops down suddenly from the clouds and collects in the ravines, and the valleys into which they lead, often cause great floods, which are especially disastrous on account of their suddenness, which leaves little or no time for escape. The following is a newspaper account of cloud-bursts which occurred near Fort Keogh, Mont., a few

years ago, and which are merely samples of similar occurrences which frequently take place in Montana and adjacent territories.

"Word has been received from Simmons' sheep corral, on the American fork of the Mussel shoal, that a cloud-burst there Monday evening destroyed 800 head of sheep. The cloud exploded at the head of Dry Run Creek and came pouring down in a solid wall 32 feet high, carrying off nearly the entire herd, and almost drowning a herder. The carcasses of the animals are strewn along the river for a distance of 16 miles below the scene of the disaster. The upper Yellowstone valley was visited yesterday by a terrific hail-storm, which rooted up and destroyed every growing thing in a strip of country six miles wide. Near Merill occurred a cloud hail-burst. For half an hour the hail was beyond description. There were drifts of hail 14 inches deep in some places. There was little rain accompanying the hail—simply one sheet of hail came pouring down."

The Signal Service observer at Fort Elliott, Tex., reports:[13]

"A thunder-storm began at 4.10 P.M. and ended at 7.40 P.M., moving from southwest to northeast. Hail began at 5.18 P.M. and ended at 5.26 P.M., the hailstones being spheroidical in shape and about two inches in diameter; formation, solid snow. The 'breaks' (hills) at the foot of the plains several miles northwest of station were absolutely white with hailstones for three hours after the storm. This was observed by everybody at the station; on the morning of the 26th I walked down to the Sweetwater Creek, three fourths of a mile distant, and saw great banks of hailstones which had been washed down during the night. The bottoms along the Sweetwater were literally covered with banks of hailstones from six to eight feet in depth. It was estimated that there was enough hail to cover ten acres to a depth of six feet. The hailstones killed five horses which were out on the prairie on a ranch six miles north of station. The Sweetwater Creek was higher than ever known before, the freshet destroying nearly the entire post garden. The high water is supposed to have been caused by a 'cloud-burst' at or near the foot of the plains, where the Sweetwater has its source; there was only 0.36 inch of rainfall at the station. On Sunday, May 27th, hailstones were collected on the banks of the Sweetwater, which had been washed down and lay in drifts 6 feet deep, actual measurement by the observer."

280. From these accounts and many other similar ones which might be cited, it is evident that the water in such cases does not fall as rain, but is poured down in streams; and this is especially evident from the observed effects of the Hollidays-

burg tornado, by which great holes and basins were made in the ground, and the lightest materials on the margins of these on the upper sides were not washed away; and the reason of this evidently is that the accumulated water in the cloud is poured down in streams through the ascending current of air, which at the time is too strong to allow even the largest drops to fall as rain.

It is seen from a reference to Table VII that a rain-drop of 0.2 inch in diameter at an altitude of one mile cannot fall through an ascending current of air with a velocity of 25 miles per hour; and those with still smaller diameters are carried up to where the velocity and density are such that they are supported by the current, and begin to fall only after they become so large, or the velocity of the ascending current becomes so much diminished, that they are no longer supported in the air, and fall as rain. Hence with an ascending velocity of 25 miles per hour, no rain in drops of ordinary size of 0.2 inch or less can fall directly back to the earth; but in the case of a tornado it can be carried out above, and fall at a distance from the centre, provided the drops are not too large to be carried up above the level which separates the lateral inflowing current below from the outflowing current above. The ascending velocity may be such as to carry out above the smaller drops only, and those with larger diameters are supported in the cloud in the central part of the tornado until, by blending with one another, they become much larger, or, in the case of very rapid ascending currents, until they collect into large masses which force their way down in streams of water.

281. All that has been stated with regard to immense rainfalls in cloud-bursts is applicable, with a few modifications, to the sudden falls of immense quantities of hail. In this case, however, the circumstances must be such that the out-turning current is at a great altitude and the strength of the ascending current is so great that, instead of the accumulation being down very far below the region of freezing, it must be higher up, but not necessarily in this region. For the hail-stones, formed in the manner described, may fall directly back from

above, or be carried out above and come around and be carried up into this place where the ascending current is too strong to allow them to fall, and not strong enough to carry them up where the current above would carry them out where they can fall, and so they remain there a considerable time without much loss from melting even when collected below the region of freezing temperature, until from some cause the energy of the tornado becomes exhausted suddenly, or the accumulation becomes too great to be supported, and having once broken through in spots, it collects together into streams, as rain does in a cloud-burst, and falls to the earth in a very short time, leaving a great depth of hail over a short and very narrow district of country.

If the collection of hail is up in the freezing-region, then, by the continual freezing of the rain carried up from below and coming in contact with the hail-stones, they grow until they become so large, or until the velocity of the current is so much diminished, that they can be no longer supported in the air. This may not be until they become very large. Accordingly, hail-stones of enormous size are sometimes observed, and even great chunks of ice, seemingly formed by the freezing together of the water and the hail-stones collected in the vortex of the tornado (§ 276).

FAIR-WEATHER WHIRLWINDS AND WHITE SQUALLS.

282. These are simply small tornadoes which occur during fair weather whenever and wherever the proper conditions are found, which are generally when there is an unusually rapid decrease of temperature with increase of altitude, or else a very nearly saturated atmosphere. They are sometimes observed on land, but mostly on lakes and at sea, and may be accompanied by a water-spout or not, according to the violence of the gyratory motion and the amount of aqueous vapor in the atmosphere. At first a small cloud is formed in the clear sky, which gradually increases from the condensation of vapor rapidly carried up in the central part of the gyrating and

ascending air just as in the case of any other tornado. The gyrations may sometimes continue to increase in violence and extent, and the cloud to spread until it assumes the dimensions of a tornado such as usually takes place in a cyclone; but it is mostly confined to only a very narrow streak, and its violence, although very great at the centre, is not felt at a short distance away. Although the gyrations are violent very near the centre, yet decreasing in velocity, at least a little above the earth's surface where the friction is small, somewhat according to the law expressed in § 42, at a short distance from the centre they are entirely harmless, and they may pass very near without doing any injury. If, however, they pass directly over houses on land or a ship at sea, their destructive effects may be very great and sudden.

When these small tornadoes are accompanied by a water-spout and are first seen at sea at a distance, they can generally be avoided, or, if not, preparations can be made to guard against danger; but where they are suddenly formed on the spot and drop down, as it were, from above, they give no fore-warning of approaching danger. A remarkable example of this kind was related in the New York *Herald* of December 10, 1878. The British bark *Bel Stuart*, Captain Harper, on the evening of November 14, 1878, 160 miles from Cape Sable,

"was struck by a white squall in a comparatively smooth sea and clear sky, which swept her decks and created consternation on board. At 6 P.M. of the same day, all hands being on deck after supper, a strange sighing of the wind was observed by the watch, and the sky became suddenly threatening without corresponding indication of the barometer, which showed a rising tendency; Captain Harper and his first officer were on the deck at the time. All hands noticed the peculiar change in sea and sky, and were discussing it, when, without a moment's notice, the sea forward seemed to swell up to meet the lowering sky, and swept the bark across her bows, carrying away her foretop-gallant-mast, jib, jib-boom, foretop-mast stays, and the maintop-gallant-mast, with all their accompanying sails. In a moment, as it seemed, the bark, with all sail set, in a fair wind, with a moderate sea, was left a comparative wreck to wallow in the trough of the tremendous seas which had followed the spiral volume of water. Two minutes before the fatal catastrophe, Captain Harper says, there was no indication of the water-spout."

It seems from this account that the bark ran into the spout as it was being formed and before it became visible. The sighing of the wind was caused by the rapid gyrations of the air, and the threatening sky by the incipient condensation of the aqueous vapors carried up by the ascending current. The barometer, before its very near approach, remained unaffected, because, as we have seen, the barometric pressure is diminished only at and very near the centre, and it was by this that the uprising of the sea was caused which swept across the bark.

283. When we have the conditions of a water-spout in fair weather with little moisture in the air, we have what are called *white squalls*, or fair-weather whirlwinds. In such cases the dew-point is so low, and consequently the cloud, when formed, is so high, that the gyrations may not be able to bring it down, in the form of a spout, to the sea. But still the gyrations and the rapidly ascending current in the central part are there, and the rising and boiling of the sea below. High up in the air also, directly over the boiling of the sea, is a patch of white cloud, formed by the condensation of the vapor in the ascending current when it arrives at the height at which it begins to be condensed. This cloud may eventually extend over a considerable portion of the heavens, but at first it is a small cloud in a clear sky, as represented in Fig. 12, and is white because of the great amount of reflected light. It has no doubt the funnel shape beneath, but being high up, and the observer being generally nearly under it, this feature is hardly ever observed. If the

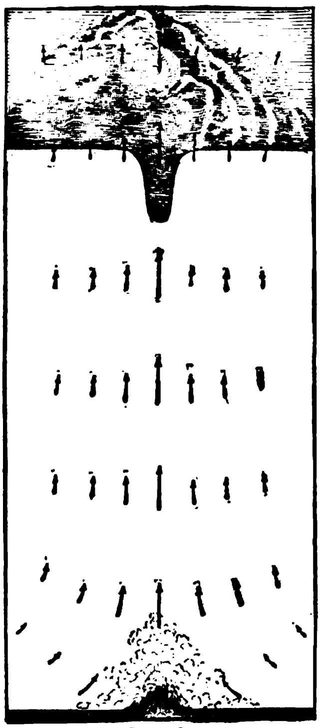


Fig. 12.

air be not too dry, it may be followed by a shower of rain.

Peltier says:

"White squalls are very rare, but they are sometimes met with between the tropics, especially near elevated lands. They are generally violent and of short duration. They often take place when the sky is clear, and without any atmospheric circumstances giving notice of their approach. The only thing which indicates their proximity is the boiling of the sea, which is very much agitated by the violence of the winds. Many of these squalls, which commence by a little cloud, or even without any visible cloud, are soon accompanied by violent rains and thick clouds."

On the west coast of Africa these little tornadoes or whirlwinds are called *bull's-eye squalls*. According to Piddington, the Portuguese describe such a squall as " first appearing like a bright white spot at or near the zenith, in a perfectly clear sky and fine weather, and which, rapidly descending, brings with it a furious white squall or tornado."

From the preceding description it is evident that these squalls differ but little in their nature from the small tornadoes and water-spouts met with in higher latitudes, such as the one which nearly wrecked the bark *Bel Stuart*, except that the atmosphere is usually too dry for the formation of a water-spout.

WHERE TORNADOES ARE MOST LIKELY TO OCCUR.

284. It has been shown that there are two principal conditions upon which tornadoes depend, and that in the absence of either of these they cannot take place. The one is the state of unstable equilibrium of the air, and the other a gyratory motion with reference to any assumed centre. It is not necessary that the centre shall be stationary, but simply that the motion of the air around it shall be such that, when it is drawn in toward this centre, it shall run into a gyration around it. When we have these two principal conditions, the other, that that there shall be some slight initial disturbance to cause the air to burst up at some point through the strata above, can scarcely ever be wanting. The places and times, then, in which

these two principal conditions are found are those in which tornadoes are most likely to occur. Of these two, however, the unstable equilibrium is the most important, since it more rarely occurs than the other, which is scarcely ever so entirely absent as not to give at least some gyratory motion which becomes violent very near the centre.

285. With regard to fixed areas on the earth's surface where the unstable state is most readily induced at all seasons of the year, these are found where, in the general motions of the atmosphere as deflected by continents and mountain ranges, currents of air at the earth's surface which come from a warmer latitude, or at sea from a much warmer continent, are caused to flow under the cold upper strata where the normal motion is nearly eastward, and where consequently the temperature is the normal one, not affected by such motion as takes place in the lower strata. One such place is found in the Mississippi Valley, and especially between the Mississippi and the Rocky Mountain range, where the air currents of the lower strata are from lower latitudes, comprising the Gulf of Mexico, curving around first northward and then more eastwardly under the higher upper strata which pass over the top of the range of the Rocky Mountains, and directly or nearly eastward without having their temperatures changed from the normal temperature of the latitude by such deflections. And this is especially the case in the summer season, when the interior of the continent is warmed up and the air of the lower strata is drawn from lower latitudes far up into the higher latitudes on the eastern side of the Rocky Mountains, and the isothermal curve there is deflected very far toward the north. From this cause the temperature of the lower strata of this region becomes unusually great relatively to that of the strata above; and if the complete unstable state is not induced from this alone, it is readily brought about by the addition of any small effect from some other cause, as from extremely warm weather in which the earth's surface and the lower air strata become abnormally heated. The great moisture of the air in these southerly winds is also favorable to the induction of the un-

stable state, since this state is more readily brought about in air nearly or quite saturated.

286. The other condition of tornadoes, that of a relative gyratory motion with regard to any point, is also found to an unusual extent in this region, especially in the winter season. For the southerly current curving around toward the east causes a pressure toward the right, giving rise to the permanent barometric gradient of increasing pressure toward the central part of the permanent area of high pressure in the Atlantic Ocean east of Florida; and on account of this there is a counter-current between this and the Rocky Mountains, flowing down toward Texas, just as in the Atlantic Ocean the Greenland current flows down to Florida between the Gulf Stream and the coast of the United States. There is evidence of such a current in this region in the averages of all seasons since the resultant of these has a large southerly component, and especially is this seen in the isotherms of the Mississippi Valley extending in a somewhat northeasterly and southwesterly direction, so that the mean temperature of New Mexico and the northern part of Texas is the same as that of places east of the Mississippi from five to ten degrees farther north. In the summer season this flow of cold air down toward the Gulf is confined to a comparatively narrow belt close to the mountain range; for then the warm currents from the Gulf are drawn farther up toward the northwest. At this season the northern part of Texas has the same mean temperature as Minnesota, and the isotherms are nearly north and south in direction, and the contrast between the temperature of the warm southerly winds on the one side, and the colder northerly ones on the other side, is similar to that of the cold wall between the Gulf Stream and Greenland current in the Atlantic Ocean. This tendency of the air, therefore, to flow in contrary directions gives the second condition of a tornado to a greater degree in this region than in almost any other. Both of the two principal conditions of tornadoes, therefore, are found in a preeminent degree in the Mississippi Valley, and especially between the Mississippi and the Rocky Mountains. Hence there

is perhaps no part of the world where they prevail more than in this region, and especially in that part of it in the middle latitudes west of the Mississippi River embracing Kansas and Missouri.

Near the Rocky Mountain range, and some distance to the east of it, the conditions are not so favorable, both on account of the dryness and the lower temperature of the air. Hence in this region large destructive tornadoes do not prevail much, though small tornadoes with water-spouts are of frequent occurrence in the lower strata of the atmosphere.

287. Another place on the globe where the normal condition of the atmosphere approximates to the unstable state is on the west coast of Africa and extending a considerable distance westward over the ocean. Here the warm trade-winds from the northern part of Africa run under the comparatively very cold air of the upper strata, moving eastward over the Atlantic, and thus causes a very rapid decrease of temperature with increase of altitude, very nearly, if not quite equal to that which produces the unstable state in unsaturated air. But the excessive dryness of the air here, coming from the coast of Africa, is a condition which is not favorable to the formation of large tornadoes accompanied by water-spouts, but very small tornadoes or whirlwinds without spouts, usually called white squalls, are very frequently seen in this region and are here called bull's-eye squalls.

288. For well-known reasons, the unstable state occurs mostly in the cloud region, and hence the tornadic gyrations usually commence there first, and are afterwards propagated downward to the earth's surface. The unstable state, however, is often produced in the lower, unsaturated strata of the atmosphere, even when very dry; and then if the other condition of a tornado is present, small tornadoes at least may be formed, even where this unstable state does not extend to the upper strata, and thus tornadoes occur not only in cyclones, but elsewhere and in clear weather. In such cases, however, the effects do not reach very high into the atmosphere.

The unstable state in unsaturated air occurs mostly on very

dry and sandy soils, with little heat conductivity, when the weather is very warm and the heat rays of the sun are unobstructed by any clouds above. The heat thus accumulates in the surface strata of the soil and the lower strata of the atmosphere, and thus is brought about the unstable state, at least up to a low altitude, even in clear and dry weather.

The same is often found, also, in very calm weather over the surface of seas and lakes. The surface of the water becomes heated, and also the lower strata of the atmosphere, by heat rays passing directly down and by those reflected back until the unstable is brought about.

289. The season of the year in which tornadoes mostly occur is that in which the atmosphere in its normal state for the season approaches most nearly the unstable state. This seems to be in all parts of the world in the summer season, and in the United States at least in the early part of summer, May or June. Hence, although tornadoes occur at all seasons and in every month of the year, yet, according to Finley's researches,[73] "summer is the season of greatest frequency," and "June is the month in which they occur most frequently." The relative frequency of their occurrence in the United States during the winter, spring, summer, and fall seasons was found to be respectively as the numbers 35, 215, 240, and 86. These numbers indicate a great difference between summer and winter in the frequency of their occurrence, and also, since the numbers for spring and summer do not differ much, that the maximum occurs very early in the summer.

290. For the very same reason that tornadoes occur mostly during the warmest season of the year and very rarely in the winter season, they should occur during the warmest part of the day and seldom at night. For during the day the surface of the earth and the lower strata become very much warmed up, and at night the reverse takes place from nocturnal cooling, while the temperature at a moderate elevation is subject to only a small diurnal variation. Hence, when the general state of the atmosphere, aside from its diurnal variation, is very nearly that of the unstable state, this state is frequently induced

during the warmest part of the day by this diurnal and other abnormal variations, but very rarely at night. Accordingly Finley found that "tornadoes are most frequent in the afternoon between noon and 6 o'clock," and that "the hour during which the greatest number of tornadoes occurred was from 5 to 6 P.M.," and that "the next hour was from 4 to 5 P.M."

The same is true with regard to small tornadoes and water-spouts, as well as those which occur in cyclones. M. Defranc" remarks that he "never saw a water-spout before 10 o'clock in the morning nor after 5 o'clock in the evening," that "they never appear during the night nor during the winter, and that there are always two circumstances attending them: the first is the presence of the sun during or a little before the phenomenon; the second is the absence of the wind, or only a very feeble one, except in the space occupied by the water-spout."

This refers mostly to small water-spouts on seas and lakes, depending upon the conditions referred to in a preceding paragraph.

291. Tornadoes occur mostly in the summer season and during the warmest part of the day, not only because the vertical gradient of decreasing temperature is greatest at these times, but also because a smaller gradient is required to induce the unstable state then than during the coldest season of the year and the coldest part of the day. By referring to Table III, Appendix, it is seen that with a hot surface temperature of the air during the warmest part of the day in the summer season, say 35°, the unstable state for saturated air is induced with a vertical gradient of decreasing temperature of 0.35° for each 100 meters, while, if the temperature were that of freezing, the gradient would have to be 0.63° for each 100 meters, and hence nearly twice as great as in the former case. So great a gradient as the latter is rarely if ever found in winter, even during the warmest part of the day.

In the case of small fair-weather tornadoes and water-spouts, such as are observed on seas and lakes, unless the atmosphere is very near the point of saturation, the vertical

temperature gradient required is very nearly that of dry air, which is never found in the winter season, and only during the hottest part of the day in summer. Hence these are never observed at night even in the summer season.

292. Where the unstable state has been very nearly induced from some other cause, as a very hot surface temperature during a warm, clear day, it may often be consummated by the burning of dry brush on the earth's surface, or of a cane-brake, or by a large fire of any sort, and thus great whirlwinds, and even showers of rain accompanied with thunder, may be produced. The vertical columns often visible in such cases are composed of dark smoke brought in from all sides, and carried up in the centre. These assume all the usual forms of water-spouts, and of course often contain also condensed vapor in the form of a water-spout, concealed by the smoke, else rain would not be produced.

Mr. Olmsted[67] has given an interesting account of such phenomena arising from the burning of a cane-brake on the shores of the Black Warrior, in Alabama. Columns of smoke of various forms were witnessed, which assumed the usual forms of water-spouts, some extending up to only a moderate height with a funnel shape at the top, while others were very slender columns extending up 300 yards into the clouds of smoke, and all were accompanied with a whirling motion of the air and the smoke. Both Redfield[68] and Espy[61] have given a number of statements made by eye witnesses of the effects of great fires in producing whirlwinds, rain, and thunder, from which it appears evident that they are at times followed by a considerable amount of rain. And it was proposed by the latter to try the experiment of producing artificial rains in time of drought by burning great quantities of brush-wood, or by means of prairie fires in the West when the grass is dry.

SAND-SPOUTS AND DUST WHIRLWINDS.

293. In very hot, dry climates, where there is a sandy soil, sand-spouts and dust whirlwinds are of frequent occurrence.

The dry air of such climates, especially over a sandy soil, is often in a state of unstable equilibrium from the accumulation of heat on the earth's surface, for a sandy, dry soil conducts it very slowly down into the earth, and there are then generally all the conditions of a whirlwind and a water-spout, except the vapor in the air to condense, for the condition of an initial whirl of the air can scarcely ever be wanting where there is not a perfect calm. The gyrations of the air and the ascending currents are the same as in a water-spout, but instead of aqueous vapor, sand or dust collected and drawn in from the vicinity is carried up. The inflowing and spiral currents from all sides towards the vortex, up to a considerable height often, keep it near the centre in the form of a column or pillar of sand, if the whirlwind is well developed over a small area only, with very rapid gyrations and a strong ascending current. The height of the column depends upon the strength of the ascending currents and the altitude at which they are turned outward from the vortex; for, as in cyclones and water-spouts, where there is a flowing of the air in from all sides below, it must flow out again above a certain altitude, depending upon the different circumstances under which the whirlwind takes place. Sand-spouts are frequently observed in Arabia, Persia, and India, and also in Arizona and other places in the western part of the United States, where the climate is very dry. In the hot, dry climate of Australia, situated in the dry zone of the southern hemisphere, these pillars of sand are said to be often more than a half-mile in height.

Humboldt, in his "Aspects of Nature," refers to the sand-spouts during the dry season on the Orinoco, South America, and says that "like conical-shaped clouds, the points of which descend to the earth, the sand rises through the rarefied air on the electrically charged centre of the whirling current, resembling the loud water-spout, dreaded by the experienced mariner" (§ 119). He seems to think, however, that electricity has something to do with the phenomena.

Where the whirls take place over a considerable area, and do not become concentrated into rapid gyrations near the

centre, and the ascending currents do not extend up very high, they give rise to dust whirlwinds, which often overtake caravans and travellers in the deserts; and if they produce no fatal effects, they are at least very unpleasant and annoying. These are very frequent in India and the Sahara, or Great Desert of northern Africa, and in fact in all places in the warmer latitudes where there is a dry, sandy soil.

When the air is nearly calm, the sand and a thin stratum of atmosphere in contact with it become heated very much above the ordinary temperature of the air a little above the surface. The inflowing currents of the whirlwind from all sides collect this warm surface stratum into the central part of the whirlwind and cause the whole interior to be of an extraordinarily high temperature. The much-dreaded simoom, stripped of all exaggerations, is most probably simply one of these dust whirlwinds.

294. Sand-spouts, as well as water-spouts, have been observed to be hollow. Of a whirlwind observed at Schell City, Mo., in the summer of 1879, it was said:[89]

"There were no suface winds strong enough to bear dust along the surface of the ground, but the dust carried up in the vortex was collected only at the vortex of the whirl. The dust column was about two hundred feet high, and perhaps about thirty or forty feet in diameter at the top. The direction of rotation was the same as of storms in the northern hemisphere. Leaving the road, the whirl passed out on the prairie, immediately filling the air with hay, which was carried up in somewhat wider spirals, the diameter of the cone thus filled with hay being about one hundred and fifty feet at the top. It was then observed also that the column was *hollow*. Standing nearly under it, the bottom of the dust column appeared like an annulus of dust surrounding a circular area of perfectly clear air. The area grew larger as the dust was raised higher, being about fifteen or twenty feet wide when last observed."

The sand-spout and the dust whirlwind, where well developed and concentrated, are free from sand or dust in the centre, for the same reason that the water-spout is free from cloud or condensed vapor. The centrifugal force of the gyrations keeps it off at a distance where this force is just equal to that of the indrawing currents which tend to drive it in toward the vortex.

BLASTS OF WIND AND OSCILLATIONS OF THE WIND-VANE.

295. The wind is often observed to blow in blasts, with an oscillating vane and unsteady barometer. This arises from the air running into numerous whirls, or gyrations, while it at the same time has a progressive motion. As in a cyclone, while passing over any place, the wind is first from one direction and then, in the course of a day or two, gradually veers around to another, often to one in nearly a contrary direction, so, in the passages of small whirlwinds, the vane in like manner oscillates from one direction to another in a few minutes, the manner and range of oscillation, as in the case of a cyclone, depending upon which side of the vane the centre of the whirl passes, and the distance from it. If the centre of the whirl passes over the vane, then there is a very sudden oscillation of the vane through a range of 180°, or nearly, especially where the diameter of the whirling air column is small.

As in a cyclone the velocity of the wind is the greatest on the side called the dangerous side, on which the direction of cyclonic motion coincides with that of the general progressive motion, and is comparatively small, or may entirely vanish, on the opposite side, so in one of these small whirls of air the velocity on the one side is very much increased above the average, which is that of the general progressive motion, while on the other it is much diminished, and there may be almost a calm. Wherever the air is in the unstable state these little whirls are very numerous, and consequently, in their passage over any place, they cause the wind to blow in blasts, and a very frequent oscillation of the vane from side to side occurs, sometimes from right to left, and at other times in the contrary way.

These little whirls in the atmosphere are especially liable to occur in connection with cyclones extending over a considerable area; for in these, especially up in the cloud region, the air is in the unstable state, and, on account of the gyrations of the cyclone and the general agitation of the air, the sum of the

moments of gyration, with regard to any point where there is a rushing up of the air of the lower strata through those above, can scarcely ever be 0, and hence there are both of the two principal conditions of such little whirlwinds. These may form small secondary cyclones contained within the larger, or tornadoes and water-spouts, or simply local whirlings in the atmosphere of small extent and no great violence, but sufficient to cause intermittences in the steadiness of the velocity of the wind and oscillations in its general direction. Hence there is generally great unsteadiness in the velocity and direction of the wind in a cyclone, and the greatest injuries usually arise from the violence of the wind on the side of these subsidiary whirls where the direction of motion coincides with that of the gyratory motion of the cyclone.

These blasts and oscillations of the vane are generally observed on the clearing-up side of a storm. As the central area of a cyclone is warmer than the surrounding parts, and the upper colder strata in middle and higher latitudes move eastward faster than the lower strata, the effect is to cause a more rapid decrease of temperature with increase of altitude, and hence to induce the unstable state which gives rise to these whirls. As the air is comparatively dry on this side of the storm, the small amount of vapor remaining is usually carried up in the central part of the whirl to a considerable altitude before condensation takes place, and then it forms a patch of whitish fracto-cumulus cloud of greater or less extent, or even sometimes to a large dark cloud if the extent and duration of the whirlwind are sufficiently great. As the condensed vapor is carried up in the shape of a cumulus cloud to a considerable height above its base, and the progressive velocity of the upper strata is greater than that of the lower, the tops are blown forward in the general direction of the currents, so that they often appear of an oblong shape and indicate the general direction of the currents at that altitude. Such clouds usually appear for a while and gradually vanish by the re-evaporation of the condensed vapor forming them.

"PUMPING" OF THE BAROMETER.

296. In connection with frequent changes of the velocity, and consequently force, of the wind, there is an unsteadiness of barometrical column called "pumping," where the barometer is placed on one side or the other of a barrier to the progress of the air. This effect is given in millimeters of the barometer by the formula of § 235. According to this, where the barometer is placed against a wall or post where the wind blows normally against its surface with a velocity of 10 meters per second, the height of the barometer is increased 0.5 mm. If the velocity is increased to 20 m. per second, it becomes 2.0 mm., and hence a change of 1.5 mm. with a change of velocity from 10 m. to 20 m. per second. With a change, however, from 20 m. to 30 m. per second the change in barometric pressure would be 2.5 mm. Hence the same change of velocity where the velocity is already great gives a much greater effect upon the barometer than the same change in velocity where the velocity as yet is small. Hence in cyclones where the general velocity is great, small changes in velocity produce a considerable effect on the barometer; and it is in cyclones, therefore, where this effect is mostly observed. Much depends upon the position of the barometer. If placed in the open air, little or no effect is observed, since there is little obstruction to the wind. If placed where wind blows obliquely against the face of the barrier, the value of ΔP in the formula must be multiplied by $\cos^2 i$, i being the angle of incidence of the wind, and hence the effect is diminished proportionally. On the lee side of a barrier there is a slight depression of the barometer with the increase of velocity in the blasts arising from the dragging away of the air on that side through friction. A barometer placed in a tight room, of course, cannot be much affected, and perhaps not sensibly in any room with doors and windows closed, especially when the blasts are sudden and of so short duration that there is not time for the increase of pressure to be felt inside.

In the hurricanes of the Antilles, observation shows that these small oscillations of the barometer are closely connected with and dependent upon the blasts of the wind, and that oscillations of the vane always accompany the blasts, showing that the latter are due to small gyrations of the air. Padre Viñes says:

> "Under the influence of the blasts, the barometric column is so agitated and so irregular that it renders the reading of it very difficult, since it is scarcely possible to take an exact medium. The amplitude of the oscillations is usually from four to eight tenths of a millimeter, and sometimes more. The agitation is fitful and violent, just as the impulses which the anemometer receives and the oscillations made by the vane."

CHAPTER VIII.

THUNDER-STORMS.

297. THE greater part of violent tornadoes are accompanied by thunder, and so are properly called thunder-storms. According to Finley, "of 473 cases in which the atmospheric conditions preceding tornadoes were observed, 410 were reported as violent thunder-storms." But usually in thunder-storms there is little of either cyclonic or tornadic violence, and the wind accompanying them, when strong, is more of the character of a sudden and violent squall, though it often amounts to nothing more than a gentle wind blowing out beneath the thunder-cloud.

Much attention has been given of late years to the study of thunder-storms, both in the principal countries of Europe, and by the Signal Service and the New England Meteorological Society of our country. The principal results obtained in these studies, from very numerous observations by M. Fron in France, Ferrari in Italy, Von Bezold in Bavaria, Assmann in Germany, Mantel in Switzerland, Klossovsky in Russia, Elliot in India, and Mohn in Norway, have been given briefly by Professor W. M. Davis.[80] He has also given a report on the thunder-storms in New England in the summer of 1885.[81] To these contributions on the subject the writer is indebted mainly for the facts in what follows on this subject.

OBSERVED PHENOMENA.

298. Thunder-storms often appear to be small and imperfectly developed cyclones in which there is little or no sensible gyratory motion or barometric depression. Klossovsky regards them as small cyclones originating in certain segments near the

periphery of larger cyclones. According to Abercromby, thunder-storms are always connected with secondary cyclones. He says:[80]

"As surely as we see a secondary on the charts in summer, so certainly will thunder-storms occur during the day, though we cannot say in what portion of the small depression."

Again:

"As the secondary approaches any station, the wind draws more or less in toward the centre, and recovers its former direction after the depression has passed."

In central Germany also

"The wind in thunder-storms veers and backs with equal frequency; and not unfrequently reverses its course directly."[80]

In one of the local thunder-storms of July 29, 1885, in New England, there were indications of

"a tolerably distinct cyclonic motion of the winds within and around the oval rain-area, implying that the thunder-storm area possessed a gentle, spiral, rotary circulation on a small scale, as has been determined for storms of this kind in Europe."[81]

In Italy it is said:[80]

"The storms occur on the after-side of a small, faint area of low pressure; and that the pressure rises as the maximum phase of the storm approaches. . . . Sometimes the storms form at the end of a V-shaped 'sack' or enlargement on the side of a large area of low pressure."

And in Bavaria:

"On maps having the pressure shown from five to five millimeters, the areas of low pressure are but faintly indicated by curves in the isobars; with more detailed study, they appear in greater distinctness and are generally seen to be extensions of larger low-pressure areas, with gradients so faint that they cause no noticeable winds."

But the slight barometic depressions and cyclonic motions over a considerable area are often, perhaps generally, not observed, but a sudden change of temperature and pressure and a corresponding sudden wind-squall generally from some northwesterly direction. The sudden change of temperature is usually 10° to 20° or 25° F., and the corresponding changes of barometric pressure rarely amount to as much as 0.1 of an inch

(2.5 mm.), and these changes take place mostly in about ten minutes. In India the change of temperature is said to be only 10° to 12° F., while the change of pressure is from 0.8 to 0.15 of an inch, the greater part of which always occurs very suddenly. The corresponding sudden increase of the wind here often amounts to from 40 to 60 miles an hour.

In Italy, according to Ferrari,[10]

"Before the thunder-storm, the pressure and the relative humidity fall and the temperature rises in such a manner that the first two reach a minimum and the last a maximum at the moment of commencement of the thunder-storm; then the pressure and the relative humidity rise rapidly and the temperature falls, and both of the first often reach a maximum and the last a minimum by the end of the storm. The change of temperature is exactly the reverse of that of the relative humidity and of the pressure. The velocity of the wind, before the thunder-storm small or very nearly nothing, increases rapidly with the commencement of the storm, reaches its maximum at the end or shortly after, and then rapidly falls again."

It should be observed here that the changes of temperature, and the corresponding reverse changes of relative humidity, necessarily take place from the changes of capacity of the air for moisture with changes of temperature, and do not indicate any corresponding changes in the absolute amount of moisture in the air before and after the thunder-storm sets in. Although the relative humidity is greater, yet it is most probable that the absolute amount of vapor is less after than before the storm.

299. On the evening of June 7, 1885, barographic curves were obtained at Ann Arbor (Michigan) of two thunder-storms which passed over. These are described in the same words which had been used to describe similar changes during thunder-storms observed at Berlin:

"Before the outburst of the thunder-storm, the curves sank slowly, next rose steeply to a considerable height; . . . the curve then maintained itself at a level for some time, throughout which the thunder-shower or hail was wont to fall; on the cessation of rain, the atmospheric pressure sank steeply. . . ."

It is also stated in the same connection:

"Eye observations of the barometric height were also taken during the first storm at Ann Arbor, and immediately after the beginning of the

rainfall the barometer was observed to rise 0.07 of an inch in about ten minutes, then remained nearly stationary for about 20 minutes, when it began to sink. Preceding this thunder-storm (which came up from the west), the wind had been blowing pretty steadily from the west for an hour or two at the rate of 12 miles per hour; but as the thunder-storm approached, it fell to a velocity of about 5 miles and, coincident with the rise of pressure, increased to a velocity of 24 miles per hour, which it maintained for about 15 minutes, then in less than an hour decreased to a velocity of about 3 miles, having shifted in direction from the west to the east, but soon rose again to a velocity of 6 miles. The phenomena connected with the second storm were very similar, except that the wind fell from its highest velocity of about 30 miles to an almost calm within ten minutes; but soon rose again to a velocity of nine miles from the S. E., which continued for several hours. These changes of wind velocity are in accordance with the supposition that there was an indraught of air toward the cloud in front and rear of the storms, but that immediately under the storm there was an outward movement in every direction, the different effects at the earth's surface in front and rear of the storm being due to the movement of the thunder-storm along the earth's surface." [81]

800. According to observation, thunder-storms often appear in groups, all progressing eastwardly, in parallel directions, some abreast, and others following after. Hence there is often a succession of such storms at the same place within a short period of time, as, the same afternoon. Thus on July 29, 1885, five separate storms were traced in the southern part of New Hampshire and the eastern part of Massachusetts. Of the last and most extensive of these it is said:

"It does not seem to have been a well-united storm, but consisted of numerous loosely connected parts, from which showers of varying strength fell." [81]

On July 3 there was a group of thunder-storms in central and southern Massachusetts and central and south New Hampshire, of which no less than ten could be traced from the observations. There may have been a hundred different centres of action around which rain and thunder occurred over progressive areas, of greater or less extent and of various times of duration, from which no reports were received. In such a thunder-storm area there may be such a blending of the different storms that there are few places where rain does not fall and

thunder is not heard, and so, from reports received from a limited number of stations, it may seem to be one homogeneous storm area. But even the reports often indicate that this is not the case, as is seen above in the extract from the report of the storm of July 29, but that it consists of numerous loosely connected parts. In all such cases the relation of each distinct storm area to the general storm area is the same as that of heavy showers to the general rain area on a larger scale, § 209, where the atmosphere over a large area is in the unstable state, but in which the conditions have not been such as to give rise to a cyclone. It is simply one of the numerous places where ascending currents are for some reason started more than at other places, which give rise to rain and generally thunder, but in which the condition is generally absent which gives rise to rotary circulation.

The general form of the rain area at any given time in a compact group of simultaneous storms, and even where it may be regarded as one solid storm, is no doubt very different in different cases, and in the same storm different at different times. In the case of the storm in New England, July 21, 1885, the average form seems to have been that of the annexed figure, as obtained from a composite portrait of the forms at

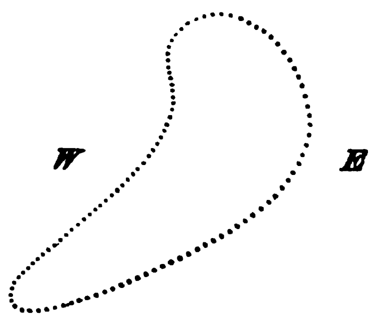


Fig. 1.

different times.[n] This also seems to be the form, according to Dr. Hinrichs, of the front convex part of the storms and squalls of Iowa.

"In northeastern Iowa the storm-front has a tendency to bend up, so as to make the squall below more nearly from the west. In a like manner in southwestern Iowa its front bends westward and hence blows more nearly from the north."

The convex front part of such storms generally seems to be well determined where the rain area assumes a somewhat regular form of any kind; but even this is often represented to be quite irregular and ill-defined.

According to Ferrari,[10] as observed in Italy,

"Thunder-storm days generally exhibit two different types of storms: in the one the thunder-storm activity is divided, and in the other one a large thunder-storm is formed, usually accompanied by others of smaller extent. In the first case small thunder-showers are formed here and there which follow after and encroach upon one another; such appear almost exclusively during the midday and afternoon hours, especially between 1 and 4 P.M. In the second case a single thunder-storm extends over a large area in a certain interval of time, while the smaller thunder-storms which appear on these days preserve their usual character almost exclusively and prefer the afternoon hours. It is sure to happen that the more extended storm on such a day is composed of more than one; but at all events, even in this case, a thunder-storm takes place which we, on account of its character, call the 'principal storm.' Such a principal storm is often the last to occur, and that which properly ends the thunder-storm activity of the period; in other cases it is, on the other hand, followed by smaller thunder-storms."

301. The antecedent and following observed phenomena are usually somewhat as follows, as deduced from a graphic average by means of a composite portrait in the case of the storm of July 9, 1885, in New England:

"An hour and a half or an hour before the storm, clouds are seen rising on the western horizon, while the winds are lightly southerly, and the temperature high (85°—95°). Nearer the storm-front, the clouds are seen to rise higher, and the temperature falls slightly, but the wind does not change significantly. The first thunder is heard from thirty to sixty minutes, and the clouds are recorded as reaching the zenith or passing overhead from ten to thirty minutes before the rain. The sudden change from gentle southerly wind to the northwest squall seldom comes more than fifteen minutes before the rain, and is generally only five to seven minutes before it; with this change the temperature falls rapidly. The squall seldom continues after the rain begins. The heaviest rain is

marked close to the rain-beginning in many cases; in others it falls from seven to twenty-five minutes later. The loudest thunder runs from ten to thirty minutes after the rain-front, and the lightning-strokes, as far as reported, fall with one exception between thirteen and twenty-seven minutes after the rain-front. Already at twenty to twenty-five minutes after the rain had begun, the western horizon is seen lighting up, and soon the clouds begin to break away; the rear edge is overhead in an hour to an hour and a half, while the rain had ceased fifteen minutes sooner on the average, its shortest duration being thirty and its longest ninety minutes. During the rain the temperature stood fifteen to twenty-five degrees lower than before the storm, and the winds were light and variable; as the storm passed over in the afternoon, an absolute rise of temperature after its passage is seldom seen, and then is faint; but a relative rise is clearly found in the maintenance of an almost uniform temperature past those hours when it ordinarily decreases most rapidly. Rainbows make their appearance between an hour and an hour and a half after the rain-beginning, and the last thunder is heard from one to two hours after the storm began." [81]

802. Although the rain-area, as represented in Fig. 1, may have considerable width, yet the energy and violence of the storm is mostly around on the front and convex edge; in fact it seems from accounts that the whole storm often assumes the form of a long extended and narrow band lying somewhat transverse to the direction of general progressive motion, but mostly, at least in New England, considerably inclined, as the southeasterly side of Fig. 1. In Italy, according to Ferrari,[82] from the few cases in which he could observe the beginning,

"the origin of a thunder-storm was a point. From this it spread out, not on all sides, but only in one direction, as is represented in Fig. 5 [here the following figure (2)]. I believe that this may generally be the way in which the thunder-storm arises. At first, also, the form of the thunder-storm is that of the sector of a circle, out of which it gradually passes into that of a band of a greater or less length."

On the 24th of June, 1888, the writer observed the first formation of a thunder-storm almost vertically over Kansas City, Mo. A little before 11 A.M. a circular dark cloud appeared a little to the north and east of his point of observation, while toward the west, and in nearly all directions, there were visible spots of clear sky. The first thunder was heard in this

cloud nearly overhead. It continued to darken and to spread at first in all directions, and soon there was an unusually heavy shower of rain, which continued nearly an hour. Before it ended the cloud seemed to cover the whole heavens evenly in all directions, and the thunder was heard on all sides. The upper clouds before and after the shower had a scarcely perceptible motion in a direction from W.S.W., and there was very little surface wind. During the rain there was a wind blowing out from the place where the cloud first formed toward the

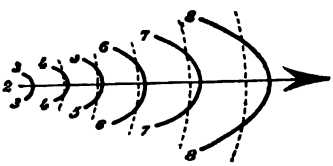


Fig. 2.

W.S.W. The breaking and thinning out of the clouds first took place in the west, and the thunder-cloud, as usual, passed slowly away in an easterly direction. The origin of this thunder-storm was evidently similar to those observed by Ferrari, but whether it continued its progressive motion a long distance and expanded laterally as it went, as represented in Fig. 2, is not known.

In Bavaria,*

"When lines are drawn so as to include the whole district over which thunder is heard at a given time, the area thus occupied is found to be a long, narrow band, at right angles to the storm's path. Storms are frequently observed that stretch from the northern limits of Bavaria to the Alps, about 200 miles, while their breadth (determined by audible thunder) is at highest 50 miles, generally about 25 miles, and often much less. In such cases the whole length of the storm-band is not measured."

Of such storms in Iowa it is said:

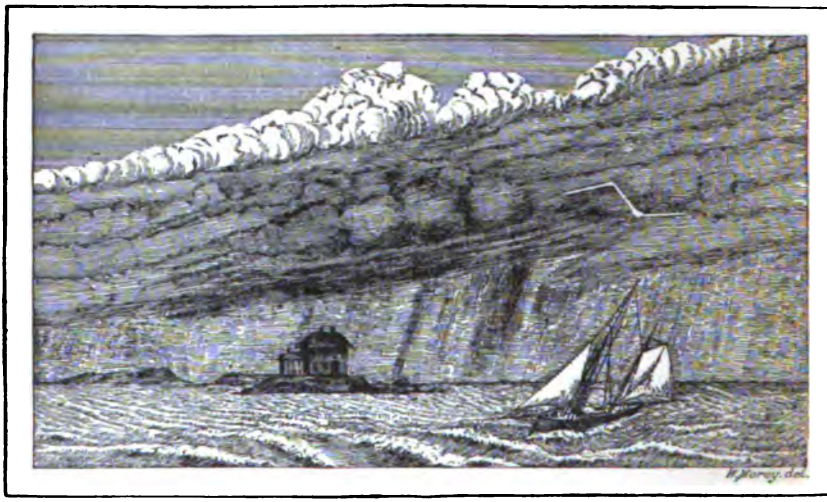
"The storm-front is fierce in its power along a considerable distance; 20 to 50 miles and more in its front along the earth are struck simultaneously. As the great storm-front sweeps on, it generally diminishes in fury, but at times it can be traced for 350 miles from the northwest to the southeast of our State."

In Switzerland at least there seem to be thunder-storms which do not have the usual easterly progressive motion, but sometimes the contrary. It is said:

"The storm occupies at first a small area, and then expands in nearly all directions, its front lines being sharply bent, almost closed, curves."

But the front lines generally seem to be similar to those represented in the preceding figure, and the great convexity on the one side undoubtedly arises from the tendency to spread faster in this direction than in any other, while in Italy and most other places the directions of spreading are confined within a small range of easterly directions, and there is a dying out of the storm in other directions.

The accompanying plate is a sketch of a thunder-storm



THUNDER-STORM OBSERVED OFF THE COAST OF MAINE, JULY, 1883.

taken by Mr. Morey off the coast of Maine, in July, 1883. The following is his account of it:

"The point of observation was about two miles out at sea.

"The storm occurred during a clear sultry afternoon, and moved from a southwesterly to a northeasterly direction, accompanied by a great display of electricity and a large amount of rain.

"The cumuli on the top of the cloud-bank were of almost snowy

whiteness, growing rapidly darker to the almost black strata forming the base.

"Previous to the disturbance the distant sky was perfectly clear except in the southwest, where a few dun-colored cumuli seemed to rest on the land. The wind was extremely light. These conditions remained after the passage of the storm, with the exception that the cumuli in the southwest had disappeared and the wind blew fresher from the northwest. Duration of storm, 30 minutes."

THE THEORY.

303. The fundamental conditions of thunder-storms, as of cyclones and tornadoes, are the state of unstable equilibrium, at least for saturated if not for dry air, and a high relative humidity. The less the latter, the more nearly must the state of the air approximate to that of the unstable state of dry air. From these conditions rapidly ascending currents and a vertical circulation arise, just as in the case of cyclones and tornadoes, and a condensation of aqueous vapor and a fall of rain or hail takes place in the ascending current, accompanied generally in summer by electrical phenomena. In what are usually called thunder-storms, the conditions are nearly or quite absent which give rise to a gyratory circulation over a large area, such as takes place in the case of cyclones, and usually the conditions are wanting which give rise to small local and violent tornadic gyrations, though, as we have seen, § 297, most tornadoes are thunder-storms. Since secondary cyclones, tornadoes, and thunder-storms are dependent upon great humidity and the unstable state of the atmosphere, it is reasonable to suppose that they must be found to exist together often in the same region having these conditions, and that there is a running together and a blending of the several forms of the storms without any distinct dividing line. Hence we find that there are thunder-storms with more or less cyclonic motions and local depressions, that they occur in regions where there are secondary cyclones, and that tornadoes are frequently observed in connection with thunder-storms. In order to the latter, it is merely necessary that, in addition to the general

conditions, there shall also be the condition, where the unstable air at any point bursts up through the strata above, which gives rise to the violent tornadic gyrations also.

We have seen that the thunder-storms seem often to arise at some given point and to enlarge and extend mostly in some easterly direction, but when the air is in the unstable state it is liable to burst up and to give rise almost simultaneously to many ascending currents and centres of local thunder-showers, which spread and eventually somewhat blend together so as to form apparently only one thunder-storm, while in reality it is composed of a number of loosely connected parts, as in the case of the thunder-storms of July 3, referred to in § 300.

Where the ascending current of one of these points of eruption is unusually strong and reaches to great altitudes, though there may not be much violent tornadic action, it gives rise to hail in the interior of the thunder-storm. Hence Ferrari states[10] "that in the region passed over by thunder-storms the hail extended in small strips in the direction of motion."

304. Let us consider first the case of a simple thunder-

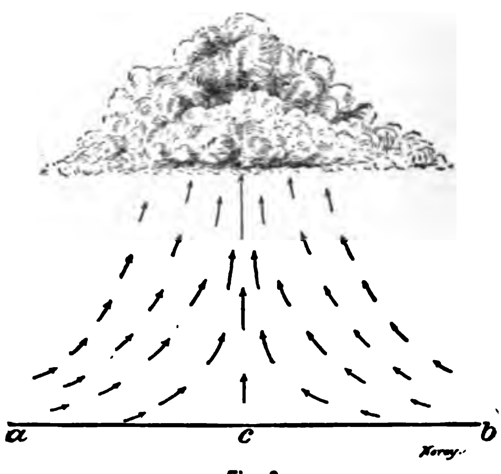


Fig. 3.

storm under conditions which are homogeneous on all sides, and in which the air has no progressive motion in any direc-

tion. If the air is in the unstable state, and over a given circular area of diameter *ab*, Fig. 3, is a little warmer and lighter than that of the surrounding parts, or for any reason receives an upward motion, there is set up, in the manner heretofore explained, a vertical circulation with an ascending current in the interior over and around the centre *c*, and an incoming current from all sides in the lower part of the air to supply the ascending current, as indicated by the arrows in the figure; while above, the current is outward in all directions, and there is a slow descent or settling down of the air on all sides. In the interior ascending current the height of incipient condensa-

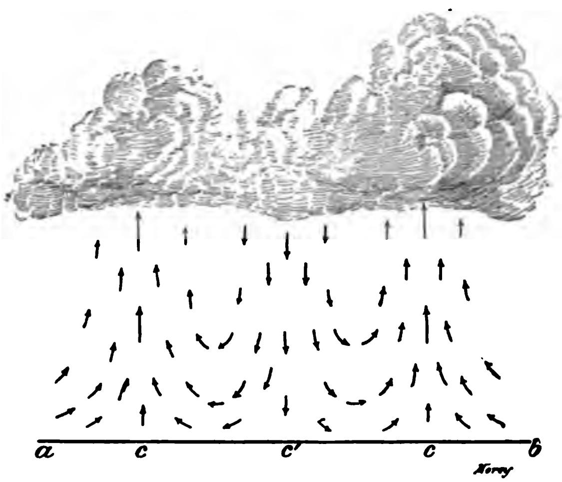


Fig. 4.

tion and of the base of the cloud depends upon the depression of the dew-point of the air, and the aqueous vapor above that height is condensed, falls as rain, and cools the air through which it falls, as already explained, until its temperature is lower than that of the surrounding air. This central cooled air, being now heavier than the surrounding air, both on account of its greater density and the amount of falling rain pressing on it, now gradually settles down and causes an outward current in all directions from the centre *c*, Fig. 4. At *c′*, Fig. 4, there is a maximum of pressure which decreases on all

sides toward c, but at some point between c' and c, but much nearer to c after the ring has expanded considerably, there is a very steep part in both the temperature and pressure gradient, exactly under the vertical between the warmer ascending current on the one side, and the colder descending current on the other, where most of the change of pressure takes place in about ten minutes generally, and where the squall exists, which encounters the inflowing current from all sides, and both are gradually retarded and deflected upward, and there is now a ring of ascending air, the middle of which is at the distance of c' from the centre. This ring of ascending air in turn becomes cooled down and changed to a descending current in precisely the same way as the first central ascending current was, and another ring of ascending air with its middle at a greater distance from the centre is developed, and so on; and this is continued as long as the air is in the unstable state. As the air on the interior side of the ring is gradually cooled down and changed to a descending current and flows under it, more of the air on the other side is thrown up into an ascending current, and the ring progresses and enlarges somewhat as a wave of water flowing out from a central point.

In Switzerland, according to observation, § 302, the conditions of this case are sometimes nearly satisfied, since the storms expand in nearly all directions and their front lines are almost closed lines. In Switzerland, in summer, the velocity of the general easterly progression of the air is small at all altitudes, and on the north side of a cyclone there may be no such motion; so that it may happen that the conditions of this case may be nearly or quite satisfied here in some cases.

305. In what immediately precedes, the air is supposed to be at rest on the earth's surface and homogeneous, or at least symmetrical, on all sides of the centre, and the area is so small at first that it is not thrown into any sensible gyration around the centre. There is consequently no reason why it should progress as a whole in any one direction rather than another, and whatever changes occur must take place equally and symmetrically in all directions. But if one side, as the south side,

is warmer and moister than the other, then the supply of energy which keeps up the vertical circulation is strongest on that side; and although the tendency is still to spread in all directions, yet it does so mostly on this side.

Again, if we suppose the whole system to be stationary, but that there is a current of air in the lower stratum next the earth's surface passing in any direction under it, as a southerly or easterly surface wind, this brings the warm moist surface air in below on the windward side, and carries the comparatively cool and dry air of the interior out to the other side, and there is, consequently, an unequal distribution of energy on the two sides, and a tendency in the ring to spread, and the whole system to move mostly, if not entirely, in the direction from which the wind and the energy come. We have a similar case if there is a wind in a great conflagration, as in a city, where the burning material is distributed equally and symmetrically in all directions. The fire extends mostly, if not entirely, in the direction from which the wind comes, because on that side is the supply of oxygen mostly to sustain the combustion, while on the other side there is air which has passed through the flame and has had its oxygen mostly burnt out.

We have a case similar to the preceding relatively, where the whole atmosphere has a regular progressive motion, say toward the east, but the part nearest the earth's surface is retarded by frictional resistance of that surface.

The more slowly moving stratum next the earth is relatively a wind blowing under the storm from the direction in which the storm, drifting in the strata above, is progressing, and consequently the effect is the same, and the tendency is for the storm not only to spread in that direction relatively to the air in which it exists, but even to subside and die out in the rear, and so to not progress in that direction. The storm then assumes the form of Fig. 1. It not only drifts in an easterly direction with the general motions of the air, but it progresses also relatively to the air, mostly in this direction, and it also spreads a little toward the south on account of the air on that side being warmer and moister, and this is especially

the case where there is a southerly wind, as there usually is in thunder-storms; but it progresses or spreads but little if any toward the north. For this reason we have the unsymmetrical form of the storm with reference to the apex and axis of progression, as represented in Fig. 1. The progress is mostly toward the east, but also toward the south, and there is a gradual dying out of the storm in the rear and on the north side, leaving a much longer wing or branch extending back on the south than on the north side, which is supported and kept up by the greater energy on that side arising from warmer and moister air, if not also from surface southerly winds, which blow somewhat under the strata above, moving more in an easterly direction.

306. Whatever the form which the thunder-storm may assume, whether that of § 304, in which there is an enlargement and a spreading in nearly all directions without much progressive motion in any direction, or that of § 305, in which there is an easterly progressive motion, and the rain and the storm areas are of the form of Fig. 1, or even more nearly a mere narrow convex band progressing as a wave and increasing in length as it goes, as represented in Fig. 2, there is a steep but short pressure gradient around the front border of the rain area as it progresses, which often gives rise to a sudden and violent squall where it passes over any place, but usually only to a gentle wind blowing out from under the thunder-cloud. This pressure gradient has been attributed to the difference of temperature within a short horizontal distance of the point c', Fig. 4, between the contiguous portions of air in which rain has not yet commenced to fall and that in which rain is falling. The rain having been produced by the condensation of vapor, at least in part, up in the high and cold strata, and then carried by the ascending current still higher, much of it is cooled down to a low temperature, and, it may be, even frozen into hail. In falling slowly through the air it cools it to a temperature considerably lower than that of the contiguous air through which rain has not yet commenced to fall, and hence the great difference of temperature and of pressure within a short distance.

The cooling of the air is also increased by the evaporation of the rain in falling through the lower unsaturated air, and also by the melting of the hail in the case of hail-fall. These seem to have been the causes assigned by Dr. Köppen in his explanation of the phenomena in the thunder-storm of August 9, 1881, which passed over Germany, so far as the writer can learn in a second-hand way from extracts from, and references to, his paper on that subject, and they are, no doubt, the true causes.

307. A difference of 1° C. in the temperature of two contiguous portions of the atmosphere extending to the top would give a difference of barometric pressure of 2.8 mm. (0.11 of an inch), which may be considered as a somewhat extreme difference, as observed in thunder-storm squalls, though of course it is undoubtedly much more in very destructive squalls, in which observations cannot be made. This, by the formula of § 235, gives at the earth's surface, where we have $P = P_0$ and T, say, equal to $T_0 + 30° = 303°$, $s = 25$ m. p. s., or about 56 miles per hour, for the velocity of the wind in the squall arising from this difference of pressure in case of no friction. Making considerable allowance for this, which is always necessary in such cases, we still have a destructive squall, and one which usually corresponds to a difference of pressure of that order. In the observation of Mr. Clayton, § 299, the difference of pressure of 0.07 of an inch (1.78 mm.) by the same formula gives $s = 19$ m. p. s., or nearly 43 miles an hour. Making a very liberal allowance for friction, we still have the greatest observed velocity of 24 miles an hour, as obtained in his observations.

We see that a difference of temperature of 1° C. extending to the top of the atmosphere would give rise to a difference of pressure from which would result a destructive squall. But thunder-storms and differences of temperature in contiguous portions of the atmosphere do not reach to the top, perhaps often not very far up. But say the observed difference at the earth's surface is 6° C., and that it gradually and uniformly diminishes with increase of altitude and vanishes at the altitude which leaves one third of the atmosphere below it. The effect upon the difference of pressure upon this supposition

would be exactly equal to that of a difference of 1° C., extending up to the top. But the observed difference of temperature in the air at the earth's surface before and after the shower is often more than 6° C., and so it is reasonable to suppose that the difference so extends up to the strata above as to give the usual observed corresponding difference of pressure.

308. There is still another cause besides the lowering of the temperature of the air which may very sensibly increase the air pressure in the region of falling rain, and that is the pressure of the rain contained in the air. This, in falling, soon acquires a maximum velocity, after which the air pressure is increased by the full statical pressure of the rain, since the whole force of gravity then upon the rain is exerted by means of friction upon the atmosphere. We cannot have generally much idea of the amount of rain and hail which may be contained in the air at any time, but we know from the amount which often falls suddenly in cloud-bursts that it is sometimes very great. If there were at any one time rain and hail falling with uniform velocity equivalent to a rainfall of 13.6 mm. in depth, it would increase the barometric pressure 1 mm., and from this alone, by the formula of § 235, would arise a squall with a velocity of about 15 m. p. s., or 34 miles per hour, making no allowance for friction.

As the falling rain, either from its cooling effect upon the air or the direct effect of its own pressure, causes a pressure gradient, and the maximum velocity of the squall is at the foot of the gradient, where the pressure is the least, the squall wind is felt a little before the arrival of the rain, since the momentum which the air has at the foot of the gradient carries it to some distance beyond before it is counteracted and the current deflected upward. Hence the beginning of the squall and the sudden change of the wind from a southerly to a northwesterly direction occurs in New England generally from five to seven minutes, sometimes considerably more, before the commencement of the rainfall (§ 301), and on land in dry and dusty weather it usually raises a great cloud of dust.

309. In the rear of the storm there is a pressure gradient,

but it is more gradual without any part which is abruptly steep, and there is consequently a flowing out of the air behind the storm, relatively to the storm itself, but no sudden squall observed, and a wind in New England thunder-storms is sometimes observed to blow out from beneath in the rear.

When the whole atmosphere has a progressive motion somewhat in the direction in which the squall blows in front, the velocity of the squall as observed on the earth is the sum of the velocity of progressive motion of the air and of the relative velocity with which the wind blows out from under the storm, and hence the observed wind for this reason alone would be much stronger in front than in the rear. As there is a gradual drawing in of the air from all sides toward the storm, or at least toward the area of ascending air whatever the form of this area, the air blowing out from under the storm encounters this and brings it to rest relatively to the storm at a greater or less distance, and at a still greater distance relatively to the earth, when the air has a progressive motion, and hence sometimes a little before the storm arrives a calm is observed. Sometimes the indrawing current is merely sufficient to reduce the progressive velocity a little, as in the case of the observations at Ann Arbor, Michigan, where the velocity of the west wind was merely reduced from 12 to 5 miles per hour and there was no perfect calm (§ 213). In this case there was a blowing out behind, or a reversal of direction, with a very small velocity.

RELATION BETWEEN THUNDER-STORMS AND CYCLONES.

310. We have seen that the origination of thunder-storms requires an unstable state of the atmosphere. If, therefore, this state is induced more readily in some parts of the cyclonic area than in others, then there must be a preponderance of thunder-storms in the former over those of the latter. But the unstable state requires a vertical gradient of rapidly decreasing temperature with increase of altitude, and this is brought about by the passage of warm southerly currents in the lower strata

of the atmosphere under colder northerly currents above, either absolutely or merely relatively. By referring to Fig. 1, § 178, it is seen that this occurs in a cyclone in about the E. S. E. octant, and in some measure in the adjacent octants on each side of this octant, in which large north and south components of the winds pass, the latter below under the former above. By referring to the opposite side of the figure it is seen that just the reverse of this takes place there, and that in the W. N. W. octant the cold northerly winds below pass under the southerly winds above, and that this is the case with regard to large components of these winds in the adjacent octants on each side. The tendency, then, of the cyclonic circulation above and below is to produce an unstable state in the E. S. E. octant, and also in some measure in the adjoining octants, but just the reverse in the opposite octants. Of course the unstable state may be brought about anywhere without any cyclonic influence, but this influence favors it in the E. S. E. and adjacent octants, and works against it in the opposite ones.

But there is also another consideration in this connection. Thunder-storms occur most frequently in the warmer and moister atmosphere of the lower latitudes than in the cooler and drier air of higher latitudes. Without, therefore, any of the cyclonic influence just referred to, there would be a maximum of thunder-storm occurrences in the southern part of a given circular area of considerable extent, and a minimum in the northern. The resultant maximum and minimum, therefore, of the two, the maxima in the E. S. E. and the S, and the minima in the opposite directions from the centre, would fall somewhere between, and most likely nearest to those arising from the cyclonic influence, and so the maximum of the resultant would fall in the S. E. octant, and the minimum in the opposite or N. W. octant.

Again it is seen from a reference to the same figure (§ 178), that in the interior of the cyclone the motion of the air below and above is somewhat in the same direction, except that it inclines in toward the centre a little, below, and out above, and so there is little or no passing, relatively, of warmer, southerly

wind below under cooler, northerly winds above, or the reverse, in any octant of the cyclone, and so the cyclonic influence which tends to produce the unstable state or the reverse in certain octants, is not found in the whole interior of the cyclone. Besides there is little tendency in this central region, on account of its smallness, and so the little difference in latitude between the northern and southern sides, toward a maximum frequency of occurrence on the south sides, or the reverse on the other.

In the outer part also of the cyclone, beyond the isobar of about 760 mm., the currents below are feeble, being under the region of high pressure, and at and near the middle of the circular calm-belt of this high pressure, and the currents above toward the outer border of the cyclone cannot be supposed to be very strong; so the cyclonic influence in this whole outer part in inducing an unstable state in any octant, or the stable state in the opposite, is very small. Besides this is a dry region, and for this reason, also, thunder-storms should rarely occur.

311. From what precedes, therefore, thunder-storm frequency should be confined, not only to the S. E. octant mostly, but also, in average cyclones, within the isobars of about 750 and 760 mm. Let us now compare these theoretical deductions with results of observations. The latter, for Russia, are contained in the following tables in percentages for the whole year, given by Klossovsky[80] from very numerous observations:

Octant	N.	N.E.	E.	S.E.	S.	S.W.	W.	N.W.
Thunder-storms	2.3	12.6	8.4	41.4	9.5	19.3	1.4	5.1
Hail-storms	—	9.4	6.2	46.9	15.6	15.6	2.1	4.2

Pressure in millimeters	735–740	740–745	745–750	750–755	755–760	760–765	765–770
Thunder-storms	0.1	0.3	5.8	37.5	48.3	7.9	0.1
Hail-storms	0.6	4.6	—	35.5	48.0	11.2	—

It is seen from the first of these tables that by far the greatest relative frequency of thunder-storms is in the S.E. octant, though there appears to be, for some reason, a slight

secondary maximum in the S.W. octant. And according to the second table, nearly all occur at the distance from the centre at which the barometric pressure ranges from 750 to 760 mm. These results accord with those deduced by Fron from the observations of thunder-storms in France, who found that they occur in the "dangerous" half of the depression between the centre and border. As the storms move mostly from S.W. to N.E. they consequently occur mostly in the S.E. octant.

Professor Hazen[83] of the Signal Service, from the study of the thunder-storms which occurred in May, 1884, in the United States, likewise found that thunder-storms generally accompany an area of low pressure, and are found in the S.E. quadrant at a distance of 400 to 500 miles from the centre.

It is seen that hail-storms and thunder-storms have about the same relative percentages in the different octants and average pressures or distances from the centres of the cyclones, and hence their relations to the cyclone centre are the same. The same is true with regard to tornadoes generally and thunder-storms. This is what we would expect from theoretical considerations, since all depend upon the same general conditions of humidity and instability of the atmosphere, and tornadoes, including hail-storms, are usually found in the thunder-storm region, and simply require the additional local condition which gives rise to tornadic gyrations and, in the case of hail-storms, to ascending currents which extend to high altitudes.

ANNUAL AND DIURNAL INEQUALITIES.

312. As thunder-storms depend upon the unstable state, and this is brought about mainly by the heating up of the earth's surface and lower strata of the atmosphere, the conditions which give rise to thunder-storms occur mostly during the warmest season of the year and the warmest part of the day. Hence thunder-storms are experienced mostly in summer and in the afternoon of the day. And where thunder-storms

and line squalls progress over the country in connection with cyclones they have been known to cease at night and commence again the next day.

On the ocean, however, the reverse of this in some measure takes place, and there seems to be a slight maximum during the winter and the night.

According to Buchan,[84] of the 23 thunder-storms which occurred at Stykkisholm in 14 years, only one occurred in any of the six warm months of the year from April to September. They occur also mostly during the night. These storms are short-lived, being in almost every case restricted to one, or at most only a few flashes of lightning and claps of thunder. Taking also the northwest stations of Scotland alone, adjacent to the ocean, the thunder-storm frequency is twice as great during the hours immediately after midnight as during the hours from noon to 4 P.M., and the storms occur here also mostly in the winter. On the east coast the reverse takes place, the diurnal maximum being in the warmer hours of the afternoon, and very few storms occur during the winter.

Along the coast of Norway also, according to Mohn, there seems to be a slight tendency to a winter maximum, not only in storm frequency, but likewise in intensity, and according to Scott, Valentia, Ireland, has a strong winter maximum of thunder-storm frequency.

The reason of contrast between ocean and land in the times of the diurnal and annual maxima of thunder-storm frequency is that the unstable state of the atmosphere upon which these storms depend is most readily induced on the former during the colder part of the day and the year, and the reverse on land. On the latter the diurnal and annual changes are great on and near the earth's surface, in comparison with what they are in the upper part of the atmosphere; while on the ocean the reverse is the case, the diurnal and annual changes being very small below in comparison with what they are above, where they are somewhat the same as over the continents. The average state of the atmosphere, therefore, on the ocean is more nearly that of the unstable state in the winter and

during the coldest part of the night than at other times, while the reverse is the case on the land. The unstable state is, therefore, more readily brought about on the ocean at these times, and hence there is the greatest frequency of thunder-storms at night and during the winter. For well-known reasons it is just the reverse of this on land.

But we have seen (§ 122) that the western sides of the continents in middle and higher latitudes partake of an oceanic, rather than of a continental, climate; and hence nocturnal and winter maxima occur, not only on the ocean, but likewise on the western sides of the continents to some distance into the interior.

The average state of the atmosphere on land is nearer to the unstable state during the summer and the warmer part of the day than it is on the ocean during the winter and the colder part of the day; and hence upon the whole there must be more thunder-storms upon land than upon the ocean, and to some extent on the eastern than on the western sides of the continents in the middle and the higher latitudes. For this reason thunder-storms and all small local storms depending upon the unstable state of the atmosphere are of greater frequency, on the average of the year, on land than on the ocean; and this is especially the case in the interiors of the continents, while small parts of the western sides, for reasons already given, are exceptions.

APPENDIX.

CONTAINING THE TABLES AND A LIST OF THE BOOKS AND PAPERS REFERRED TO IN THE PRECEDING PAGES.

TABLE I.

Gravity Correction for a Barometric Pressure of 760 mm., for each Degree of Latitude *l*, which is the Argument of the Table.

l −	Correction.		*l* +	*l* −	Correction.		*l* +
	mm.	Inches.			mm.	Inches.	
0°	1.98	0.078	90	23	1.37	0.054	67
1	1.98	.078	89	24	1.33	.052	66
2	1.97	.078	88	25	1.27	.050	65
3	1.97	.078	87	26	1.22	.048	64
4	1.96	.077	86	27	1.16	.046	63
5	1.95	.077	85	28	1.11	.044	62
6	1.94	.076	84	29	1.05	.041	61
7	1.92	.076	83	30	0.99	.039	60
8	1.90	.075	82	31	0.93	.037	59
9	1.88	.074	81	32	0.87	.034	58
10	1.86	.073	80	33	0.81	.032	57
11	1.84	.072	79	34	0.74	.029	56
12	1.81	.071	78	35	0.68	.027	55
13	1.78	.070	77	36	0.61	.024	54
14	1.75	.069	76	37	0.55	.022	53
15	1.72	.067	75	38	0.48	.019	52
16	1.68	.066	74	39	0.41	.016	51
17	1.64	.065	73	40	0.34	.014	50
18	1.60	.063	72	41	0.28	.011	49
19	1.56	.062	71	42	0.21	.008	48
20	1.52	.060	70	43	0.14	.005	47
21	1.47	.058	69	44	0.07	.003	46
22	1.42	.056	68	45	0.00	.000	45

NOTE.—The correction must be taken according to the sign placed under the argument *l*. The corrections for smaller pressures, at some altitude above sea-level, must be diminished in proportion to the pressures.

TABLE II.

The Tension of Aqueous Vapor in Saturated Air at the Temperature τ, used as an Argument.

τ	Tension.	τ	Tension.	τ	Tension.	τ	Tension.	τ	Tension.
° F.	Inches.	° F.	Inches.	° F.	Inches.	° F.	Inches.	° F.	Inches.
0	0.045	20	0.109	40	0.246	60	0.517	80	1.021
1	0.047	21	0.114	41	0.256	61	0.536	81	1.055
2	0.049	22	0.119	42	0.266	62	0.555	82	1.090
3	0.051	23	0.124	43	0.276	63	0.575	83	1.126
4	0.054	24	0.129	44	0.287	64	0.595	84	1.163
5	0.057	25	0.135	45	0.298	65	0.616	85	1.201
6	0.059	26	0.141	46	0.310	66	0.638	86	1.239
7	0.062	27	0.147	47	0.322	67	0.660	87	1.279
8	0.065	28	0.153	48	0.334	68	0.683	88	1.320
9	0.068	29	0.159	49	0.347	69	0.707	89	1.363
10	0.071	30	0.166	50	0.360	70	0.732	90	1.407
11	0.075	31	0.173	51	0.374	71	0.757	91	1.452
12	0.078	32	0.180	52	0.388	72	0.783	92	1.498
13	0.081	33	0.187	53	0.402	73	0.810	93	1.545
14	0.085	34	0.195	54	0.417	74	0.837	94	1.594
15	0.088	35	0.203	55	0.432	75	0.865	95	1.644
16	0.092	36	0.211	56	0.448	76	0.894	96	1.695
17	0.096	37	0.219	57	0.464	77	0.925	97	1.748
18	0.100	38	0.228	58	0.481	78	0.956	98	1.802
19	0.104	39	0.237	59	0.499	79	0.988	99	1.857
° C.	mm.	° C.	mm.	° C.	mm.	° C.	mm.	° C.	mm.
− 18	1.12	− 7	2.72	+ 4	6.07	+ 15	12.67	+ 26	24.96
17	1.22	6	2.93	5	6.51	16	13.51	27	26.47
16	1.32	5	3.16	6	6.97	17	14.40	28	28.07
15	1.44	4	3.41	7	7.47	18	15.33	29	29.74
14	1.56	3	3.67	8	7.99	19	16.32	30	31.51
13	1.69	2	3.95	9	8.55	20	17.36	31	33.37
12	1.84	− 1	4 25	10	9.14	21	18.47	32	35.32
11	1.99	0	4.57	11	9.77	22	19.63	33	37.37
10	2.15	+ 1	4.91	12	10.43	23	20.86	34	39.52
9	2.33	2	5.27	13	11.14	24	22.15	35	41.78
− 8	2.51	+ 3	5.66	+ 14	11.88	+ 25	23.52	+ 36	44.16

TABLE III.

The Decrease in Temperature of Ascending Air for each 100 Meters of Ascent for the different Barometric Pressures *P* and Temperatures τ, used as Arguments. Also the Weight of Aqueous Vapor in a Kilogram of Saturated Air.

P.	τ									Altitude.*
	−10°	−5°	0°	5°	10°	15°	20°	25°	30°	
mm.	°	°	°	°	°	°	°	°	°	Meters.
760	0.74	0.68	0.64	0.58	0.53	0.48	0.43	0.40	0.37	0
700	.73	.66	.63	.57	.51	.46	.42	.38	.36	660
600	.70	.63	.60	.54	.48	.43	.40	.36		1897
500	.66	.60	.56	.50	.45	.40	.37			3357
400	.62	.55	.51	.46	.41	.37				5142
300	.56	.49	.46	.42						7550
200	.48	.41	.39							10680

Weight of Aqueous Vapor in a Kilogram of Saturated Air.

mm.	gram.	gram.	gram.	gram.	gram.	gram.	gram.	gram.	gram.	
760	1.7	2.6	3.8	5.4	7.6	10.5	14.4	19.5	26.3	0
600	2.2	3.2	4.6	6.8	9.6	13.3	18.3	24.8		1897
400	3.3	4.8	7.2	10.2	14.4	20.0				5142
200	6.5	6.7								10680

* Computed with the temperatures of § 13.

TABLE IV.

The Heights, in meters, of Incipient Condensation in ascending currents of Air, for the Temperatures τ, and the Depressions of the Dew-point, $\tau - d$, in Centigrade degrees, used as arguments.

$\tau - d$	Air Temperatures τ.										
	35°	30°	25°	20°	15°	10°	5°	0°	−5°	−10°	−15°
1°	128	127	127	126	126	125	125	125	125	125	125
2	255	254	253	252	251	250	250	250	250	250	250
3	382	380	379	377	376	375	375	375	395	375	375
4	509	506	505	502	501	500	500	500	500	500	500
5	635	632	630	627	625	624	624	625	624	624	624
6	761	758	755	751	748	747	746	748	746	748	
7	886	883	879	874	871	869	868	870	868	870	
8	1012	1009	1003	997	993	991	990	991	990	992	
9	1137	1133	1127	1120	1115	1113	1111	1112	1112	1114	
10	1262	1257	1250	1242	1236	1234	1232	1232	1233	1235	
11	1387	1382	1374	1364	1357	1355	1352	1351	1352		
12	1511	1506	1497	1485	1478	1476	1472	1470	1470		
13	1635	1630	1620	1606	1599	1596	1592	1589	1588		
14	1759	1754	1743	1727	1720	1716	1711	1708	1705		
15	1883	1877	1865	1848	1841	1836	1830	1826	1822		
16	2005	1999	1985	1969	1962	1956	1950	1945			
17	2127	2120	2105	2090	2083	2076	2070	2065			
18	2249	2241	2225	2211	2203	2196	2190	2185			
19	2371	2361	2344	2332	2323	2316	2310	2305			
20	2493	2480	2463	2452	2443	2435	2430	2425			
21	2614	2599	2582	2572	2563	2554	2548				
22	2734	2718	2700	2691	2683	2672	2665				
23	2854	2837	2819	2810	2802	2790	2681				
24	2974	2956	2938	2929	2921	2908	2795				
25	3093	3075	3056	3048	3040	3025	3009				
26	3214	3192	3173	3165	3158	3141					
27	3332	3309	3290	3282	3275	3256					
28	3450	3426	3407	3399	3391	3371					
29	3568	3543	3524	3515	3506	3496					
30	3685	3660	3640	3631	3621	3600					
31	3801	3775	3757	3749	3737						
32	3917	3891	3874	3867	3854						
33	4032	4006	3990	3984	3971						
34	4147	4120	4107	4101	4088						
35	4261	4235	4223	4218	4205						
36	4375	4349	4338	4333							
37	4488	4462	4452	4448							
38	4601	4576	4566	4562							
39	4713	4690	4680	4686							
40	4826	4804	4794	4790							
41	4939	4918	4908								
42	5053	5032	5022								
43	5168	5146	5136								
44	5284	5261	5250								
45	5400	5376	5364								
46	5518	5492									
47	5638	5609									
48	5759	5727									
49	5881	5846									
50	6005	5966									

TABLE V.

Containing several functions defined and referred to in the preceding pages, useful in facilitating the computations of many of the Formulæ. The functions are given for each Fifth Degree of the Latitude l, used as an argument.

l	$\sin l$	$\cos l$	ω in Meters.	$\frac{e}{s}$	$2n \sin l$	G
0°	0.00000	1.00000	465.0	0.00000000	0.0000000	0.0000
5	.08716	0.99619	463.2	130	127	0.0137
10	.17365	.98481	458.0	258	253	.0273
15	.25882	.96593	449.2	385	377	.0406
20	.34202	.93969	436.9	509	499	.0537
25	.42262	.90631	421.5	629	616	.0664
30	.50000	.86603	402.7	744	729	.0785
35	.57358	.81915	380.7	853	837	.0901
40	.64279	.76604	356.3	956	937	.1010
45	.70711	.70711	328.8	1052	1031	.1111
50	.76604	.64279	298.9	1139	1117	.1204
55	.81915	.57358	266.7	1219	1195	.1287
60	.86603	.50000	232.5	1288	1263	.1361
65	.90631	.42262	196.6	1348	1322	.1424
70	.93969	.34202	159.0	1398	1370	.1477
75	.96593	.25882	120.4	1437	1408	.1518
80	.98481	.17365	80.8	1465	1436	.1547
85	.99619	.08716	40.6	1481	1453	.1565
90	1.00000	0.00000	00.0	0.00001487	0.0001458	0.1571

TABLE VI.

Height in Meters ΔH of a Column of Pure and Dry Air, corresponding to a Millimeter of Barometric Pressure, at different Temperatures and Pressures, Computed by the Formula $\Delta H = \frac{7993}{p} \cdot \frac{T}{T_0}$

Barometric Pressure in Millimeters.	Temperature by the Centigrade Scale.									
	− 10°	− 5°	0°	5°	10°	15°	20°	25°	30°	35°
760	10.13	10.32	10.52	10.71	10.90	11.10	11.29	11.48	11.67	11.87
750	10.27	10.46	10.66	10.85	11.05	11.24	11.44	11.63	11.83	12.02
740	10.40	10.60	10.80	11.00	11.20	11.39	11.59	11.79	11.99	12.19
730	10.55	10.75	10.95	11.15	11.35	11.55	11.75	11.95	12.15	12.35
720	10.70	10.90	11.10	11.30	11.51	11.71	11.91	12.12	13.32	12.52
710	10.85	11.05	11.26	11.67	11.68	11.88	12.08	12.29	12.49	12.70
700	11.00	11.21	11.42	11.63	11.84	12.05	12.26	12.46	12.67	12.88
690	11.16	11.37	11.58	11.80	12.01	12.22	12.43	12.64	12.86	13.07
680	11.32	11.54	11.75	11.97	12.18	12.40	12.61	12.83	13.05	13.26
670	11.49	11.71	11.93	12.15	12.37	12.58	12.80	13.02	13.24	13.46
660	11.67	11.89	12.11	12.33	12.55	12.78	13.00	13.22	13.44	13.66
650	11.85	12.07	12.30	12.52	12.75	12.97	13.20	13.42	13.65	13.87
640	12.03	12.26	12.49	12.72	12.96	13.18	13.40	13.63	13.86	14.09
630	12.22	12.45	12.69	12.92	13.15	13.38	13.61	13.85	14.08	14.31
620	12.42	12.66	12.89	13.13	13.36	13.60	13.84	14.07	14.31	14.54
610	12.62	12.86	13.10	13.34	13.58	13.82	14.06	14.30	14.54	14.78
600	12.83	13.08	13.32	13.57	13.81	14.05	14.30	14.54	14.79	15.03
590	13.05	13.30	13.55	13.80	14.04	14.29	14.54	14.79	15.04	15.28
580	13.28	13.53	13.78	14.03	14.28	14.54	14.79	15.04	15.30	15.54
570	13.51	13.77	14.02	14.28	14.54	14.79	15.05	15.31	15.56	15.82
560	13.75	14.01	14.27	14.53	14.80	15.06	15.32	15.58	15.84	16.10
550	14.01	14.28	14.54	14.81	15.07	15.34	15.60	15.87	16.13	16.40
540	14.26	14.53	14.80	15.07	15.34	15.61	15.89	16.16	16.43	16.70
530	14.53	14.80	15.08	15.36	15.63	15.91	16.18	16.46	16.74	17.01
520	14.81	15.09	15.37	15.65	15.93	16.21	16.50	16.78	17.06	17.34
510	15.10	15.39	15.67	15.96	16.25	16.53	16.82	17.11	17.39	17.68
500	15.40	15.69	15.99	16.28	16.57	16.86	17.16	17.45	17.74	18.04
490	15.71	16.01	16.31	16.61	16.91	17.21	17.51	17.81	18.10	18.40
480	16.04	16.35	16.65	16.96	17.26	17.57	17.87	18.18	18.48	18.79
470	16.38	16.70	17.01	17.32	17.63	17.94	18.25	18.56	18.87	19.18
460	16.74	17.07	17.38	17.69	18.01	18.33	18.65	18.97	19.28	19.60
450	17.11	17.44	17.76	18.09	18.41	18.74	19.06	19.39	19.71	20.04
440	17.50	17.83	18.17	18.50	18.83	19.16	19.50	19.83	20.16	20.50
430	17.91	18.25	18.59	18.93	19.27	19.61	19.95	20.29	20.63	20.97
420	18.33	18.68	19.03	19.38	19.73	20.07	20.42	20.77	21.12	21.47
410	18.78	19.14	19.50	19.85	20.21	20.57	20.92	21.28	21.64	21.99
400	19.25	19.62	19.98	20.35	20.71	21.08	21.45	21.81	22.18	22.54
390	19.74	20.12	20.50	20.87	21.24	21.62	22.00	22.37	22.75	23.12
380	20.26	20.65	21.03	21.42	21.80	22.19	22.57	22.96	23.35	23.73

NOTE.—The Barometric Pressures are the Barometer Readings corrected for varying gravity, and, strictly, they, as well as the Temperatures, belong to the middle of the column ΔH. The Carbonic Acid in the air makes ΔH a very little less, and the Aqueous Vapor a little greater, than the values above.

TABLE VII.

The approximate force of the Wind, p', upon a square foot of Normal Surface, and the diameter D of a sphere of the density of water supported in the air by an ascending current, for the several Velocities s and Barometric Pressures P, used as arguments.

Velocity s, in miles per hour.	Barometric Pressures P in Millimeters.									
	760		700		600		500		400	
	p'	D	p'	D	p'	D	p'	D	p'	D
	pounds.	inches.	pounds.	inches.	pounds.	inches.	pounds.	inches.	pounds.	inches.
5	0.1	0.01	0.1	0.01	0.1	0.01	0.0	0.01	0.0	0.01
10	0.3	0.04	0.3	0.04	0.2	0.04	0.2	0.03	0.2	0.03
15	0.3	0.10	0.6	0.10	0.6	0.09	0.5	0.07	0.4	0.06
20	1.2	0.18	1.1	0.17	1.0	0.15	0.8	0.12	0.7	0.10
25	1.9	0.27	1.8	0.26	1.5	0.23	1.3	0.19	1.1	0.16
30	2.7	0.40	2.5	0.37	2.2	0.32	1.9	0.27	1.6	0.22
35	3.7	0.54	3.5	0.50	3.0	0.44	2.6	0.37	2.1	0.31
40	4.9	0.70	4.5	0.65	4.0	0.57	3.4	0.48	2.8	0.40
45	6.2	0.89	5.7	0.83	5.0	0.72	4.2	0.62	3.5	0.51
50	7.6	1.10	7.1	1.02	6.2	0.99	5.3	0.76	4.3	0.63
55	9.2	1.33	8.6	1.24	7.1	1.11	6.4	0.88	5.2	0.75
60	11.0	1.58	10.2	1.47	8.9	1.28	7.6	1.09	6.2	0.90
65	12.9	1 86	12.0	1.73	10.4	1.51	8.9	1.28	7.3	1.06
70	15.0	2.15	13.9	2.00	12.1	1.74	10.3	1.49	8.5	1.23
75	17.2	2.47	16.0	2.30	13.9	2.00	11.8	1.71	9.8	1.41
80	19.5	2.81	18.2	2.62	15.8	2.28	13.5	1.94	11.1	1.60
85	22.0	3.17	20.5	2.95	17.8	2.57	15.2	2.18	12.5	1.80
90	24.7	3.56	23.0	3.31	20.0	2.88	17.0	2.45	14.1	2.02
95	27.6	3.97	25.6	3.70	22.4	3.22	19.0	2.74	15.7	2.26
100	30.5	4.40	28.4	4.09	24.7	3.56	21.0	3.03	17.4	2.50

These functions have been computed from the formulæ of §§ 246, 247, using the temperatures given in § 13, corresponding to the several barometric pressures. For exceptionally great velocities, these functions can be obtained from those above by taking them as the squares of the velocities.

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